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Abstract

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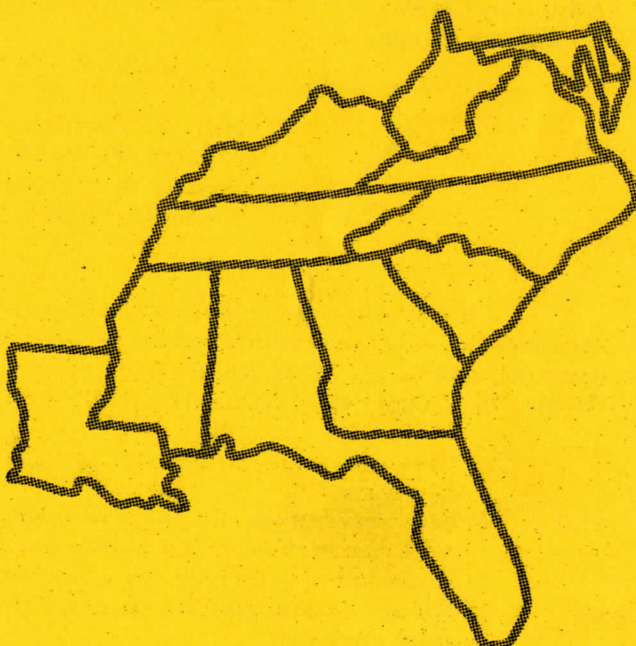

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SOUTHEASTERN GEOLOGY

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PETROGRAPHY AND DEPOSITIONAL ENVIRONMENTS OF THE MIDDLE EOCENE CASTLE HAYNE LIMESTONE, NORTH CAROLINA

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ABSTRACT

The middle Eocene Castle Hayne Limestone consists of three facies: a lower phosphate pebble biomicrudite, a bryozoan biosparrudite and a bryozoan biomicrudite. The phosphate pebble biomicrudite facies forms a discontinuous basal conglomerate. The bryozoan biosparrudite facies lies parallel to the Carolina fault and appears to be deflected basinward by the Cape Fear arch. Along the Cape Fear arch, the bryozoan biosparrudite facies thins or exists as erosional remnants. In the updip portions of the basin, the conformable transition from the bryozoan biosparrudite facies to the bryozoan biomicrudite facies is delineated by the Carolina fault. In the lower Cape Fear area, the bryozoan biomicrudite facies contains numerous diastems and is locally dolomitized.

Faunal evidence suggests that the Castle Hayne Limestone was deposited in a tropical marine biogeographic zone. The bryozoan biosparrudite facies was deposited at depths of 0-30 m, and the bryozoan biomicrudite facies at depths of 30-50 m. The distribution of the facies of the Castle Hayne Limestone suggests that during the middle Eocene: the Cape Fear arch was a positive and intermittently active feature in the lower Cape Fear area; the Carolina fault created a shelf break that directed currents onto a higher energy platform.

INTRODUCTION

Previous research has shown that units that have been mapped as lithofacies of the Castle Hayne Limestone have included units as old as Late Cretaceous and as young as early Miocene (Fallaw and Wheeler, 1963; Swift and Heron, 1969; Harris and Bottino, 1974; Wheeler and Curran, 1974; Harris, 1975; 1978; Baum, 1977; Baum et al., 1977, 1978). The purpose of this research is to present a detailed and regional petrographic analysis of the lithofacies of the Castle Hayne Limestone (as defined by Baum et al., 1978). The only previous research on the petrography of the Castle Hayne Limestone has concentrated on exposures in the Martin Marietta quarry and the Ideal Cement quarry both located in New Hanover County (Cunliffe, 1968; Upchurch, 1973; Upchurch and Textoris, 1973; Baum et al., 1978a, 1978b).

Regional Setting

The Coastal Plain has traditionally been considered a static depositional basin; however, continuing research has shown the Coastal Plain to be a dynamic system in which faulting has played a major role in the development of local depositional basins; as well as syndepositional development of lithofacies.

Within the Coastal Plain of North Carolina, three faults have been proposed: the NW-SE trending Cape Fear and Neuse faults; the NE-SW trending Carolina fault (Figure 1). These faults have been periodically active beginning in Early Cretaceous and continuing into the early Pleistocene (Harris et al., 1977, 1979; Baum et al., 1977).

Methods of Investigation

Insoluble Residue. Insoluble residue was concentrated by dissolving a dry, weighed sample (approximately 50 gm) in dilute HCl. The residue was washed several times in deionized water to remove traces of HCl. The residue was then dried and weighed. Sediment larger than silt size was retained by wet sieving on a 250 mesh sieve. The material retained on the sieve was again dried and weighed. Weight percents carbonate, sand and silt + clay sizes were calculated.

Average sand size for the sand size insoluble residue was calculated by the megascopic technique of Ingram (1965).

Petrology. Portions of the thin sections were stained for calcite using Alizarin Red S (Friedman, 1959; 1971). If dolomite or aragonite was suspected, the sample (either the chip or the thin section) was X-rayed on a Norelco-Phillips diffractometer using a copper target and nickel filter. Random or spot samples were X-rayed to determine if Mg calcite (greater than 4 mole percent) was present. No samples revealed Mg calcite (from graphs in Goldsmith and Graf, 1958 Griffin, 1971) or aragonite.

Carbonate Terminology and Classification. Folk's (1959; 1962; 1965; 1973; 1974) carbonate classification and terminology are used in this report. Porosity terminology is after Choquette and Pray (1970).

Point Count Data. A fundamental problem with point counting carbonates is that fabric selective parameters are pervasive. Thus, a single allochem can occur as original shell material (although neomorphed), or through subsequent diagenesis as moldic porosity, moldic cement, intraparticle porosity, intraparticle sediment fill or growth

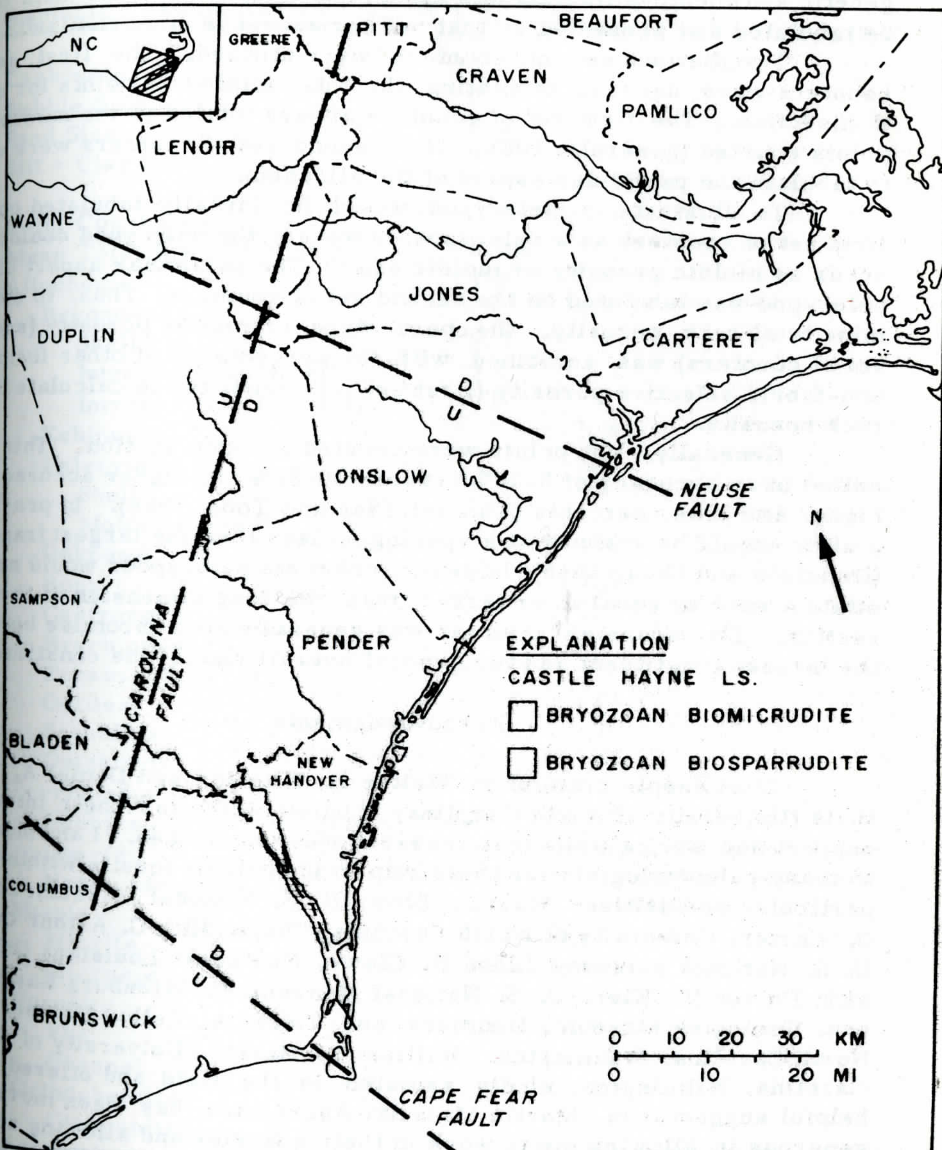


Figure 1. Geologic map of middle Eocene Castle Hayne Limestone of southeastern North Carolina (modified from Baum et al., 1978).

framework porosity; however, only one aspect of the allochem can be used in tabulating the points counted. A decision must be made on the particular aspect that is important: the original sediment or the diagenetic sediment. A method was devised in which point count data could be tabulated and presented so that neither aspect is compromised.

Two distinct sets of counters were utilized. The first set of counters were used in tabulating the total number of points for each thin section. The first set of counters always totaled to the number of points counted (generally 100). The second set of counters were used to tabulate the particular aspect of the allochem.

To illustrate, a pelecypod would be initially tabulated on the first set of counters as a pelecypod; however, the pelecypod could also occur as moldic porosity or moldic spar. The particular aspect of the pelecypod was tabulated on the second set of counters. Thus, to determine total rock porosity, the percentage of moldic porosity (second set of counters) was combined with the percentages of other forms of non-fabric selective porosity (first set of counters) to calculate total rock porosity.

Generally, 100 points were counted per thin section. This resulted in an accuracy of 8-10% between 20-80% and higher accuracy for higher and lower percents (Van der Plas and Tobi, 1966). In practice, a slide should be scanned at a spacing no less than the largest fragment (Dennison and Shea, 1966); however, complete pelecypods would necessitate a spacing equal to or larger than the long dimension of the thin section. The scanning interval was generally a compromise between the largest constituent and the general overall size of the constituents.

Acknowledgments

I am deeply grateful to Walter H. Wheeler and Daniel A. Textoris (University of North Carolina, Chapel Hill) for their interest, support and advice while this research was undertaken. I am indebted to many paleontologists for their help in identifying fossils within their particular specialties: Warren Blow, U. S. National Museum; Joseph G. Carter, University of North Carolina, Chapel Hill; G. Arthur Cooper, U. S. National Museum; Lloyd N. Glawe, Northeast Louisiana University; Porter M. Kier, U. S. National Museum; H. Wienburg Rassmussen, Geologisk Museum, Denmark; and Victor A. Zullo, University of North Carolina, Wilmington. William B. Harris, University of North Carolina, Wilmington, kindly assisted in the field and offered many helpful suggestions. Martin Marietta Aggregates has been more than generous in allowing me to work in their quarries and allowing access to their 2-inch exploratory cores. I am also indebted to Geotechnical Engineering Company for giving me 2-inch cores drilled in Brunswick and New Hanover counties.

This research was partially funded by grants from the North Carolina Department of Natural and Economic Resources, Sigma Xi,

Table 1. Mean Point Count and Insoluble Residue Data for the Phosphate Pebble Biomicrudite Facies of the Castle Hayne Limestone.

Insoluble Residue Analysis by Weight Percent (n=7)

	Mean (%)	Standard Deviation (%)
Carbonate	73.7	6.3
Pebble Size	0.1	
Sand Size	16.6	4.9
Silt + Clay Size	9.6	3.8

Point Count Data (n=9)

Mean Total Rock Porosity	6.8%	
Allochems	(%)	(%)
Bryozoans		9.1
Shell	5.7	
Intraparticle Micrite	3.5	
Intraparticle Porosity	T	
Echinoids		3.5
Pelecypods		3.2
Shell	2.2	
Moldic Porosity	1.0	
Unknown		3.0
Bone		0.7
Gastropods		0.2
Brachiopods		0.1
Foraminifera		0.1
Oolites		T
Serpulids		T
Ostracods		T
Terrigenous		
Pebbles (mostly phosphate)		32.7
Quartz		4.6
Heavy Minerals		T
Matrix		
Micrite		32.1
Microspar		T
Porosity		
Vug		4.0
Channel		1.8
Orthochemical		
Glauconite		2.5
Pyrite		2.1
	Total	99.7

the Smith Fund of the University of North Carolina at Chapel Hill, and the College of Charleston Faculty Research and Development Fund.

LITHOFACIES

The Castle Hayne Limestone consists of three facies, in ascending stratigraphic order: phosphate pebble biomicrudite; bryozoan biosparrudite and bryozoan biomicrudite.

Phosphate Pebble Biomicrudite Facies

Distribution and Occurrence. This unit consists of partially to completely phosphatized and glauconitized pebbles in a dense, well lithified, very pale orange (10 YR 8/2) to very light gray (N 8) micrite matrix (Plate 1). The pebbles range up to 40 mm ($\overline{x}$ = 12 mm) and are generally well rounded. Allochemical components include bryozoans, corals, molluscs, echinoids, sharks' teeth and bone. Since the pebbles are frequently matrix supported, the micrite is syndepositional, rather than a filtrant. Discontinuous laminae of glauconite and, to a minor extent, phosphate are frequently a conspicuous feature of this facies.

This facies forms a discontinuous basal conglomerate at the base of the Castle Hayne Limestone. It is best developed in New Hanover and Pender counties where it lies disconformably on the Cretaceous Rocky Point Member. In New Hanover County, this facies ranges from 0.0 to 1.0 m. The pebbles are scarce or absent at the disconformities with the Cretaceous Peedee Formation and the Paleocene Beaufort Formation. This is probably due to the unconsolidated nature of these formations. At all localities where this facies was observed, a diastem separates it from the overlying bryozoan biosparrudite or bryozoan biomicrudite facies.

Due to the thinness and discontinuous nature of the phosphate pebble biomicrudite facies, the areal distribution cannot be depicted on the geologic map (Figure 1). This lithofacies is assumed to underlie most of the Castle Hayne Limestone and represents a discontinuous, basal conglomerate that formed locally during the transgression of the Castle Hayne sea.

Petrology. The dominant framework components are pebbles (Table 1). When not completely phosphatized or glauconitized, the pebbles generally have nuclei that reflect the lithologies of the underlying formations. The pebbles consist of: sandy, glauconitic microsparites; sandy micrites and microsparites; sandy biomicrites and biomicrosparites; polycrystalline vein quartz; pyritized quartz arenites; bone and sharks' teeth. These pebbles occur in various stages and degrees of phosphatization and glauconitization (Plate 2); however, glauconite is more common as laminae surrounding the pebbles or replacing phosphate. Phosphate generally occurs as laminae surrounding the

Plate 1. Split core illustrating the disconformable Mesozoic—Cenozoic boundary in the Castle Hayne area: left core, bryozoan biosparrudite facies of the middle Eocene Castle Hayne Limestone disconformably on the sandy, pelecypod-mold biomicrosparrudite facies of the Upper Cretaceous Rocky Point Member of the Peedee Formation (note absence of phosphate pebble biomicrudite facies); center and right cores, phosphate pebble biomicrudite facies of the Castle Hayne Limestone on the Rocky Point Member (sacle graduated in cm).

Plate 2. Phosphate pebble biomicrudite facies of the Castle Hayne Limestone.
A. Pebbles (p) completely phosphatized (left) and partially phosphatized (right); glauconite (g) replacing phosphate (x-nichols; scale equals 1.0 mm).
B. Pebbles (p) completely phosphatized (left) and partially phosphatized (right) (x-nichols; scale equals 1.0 mm).

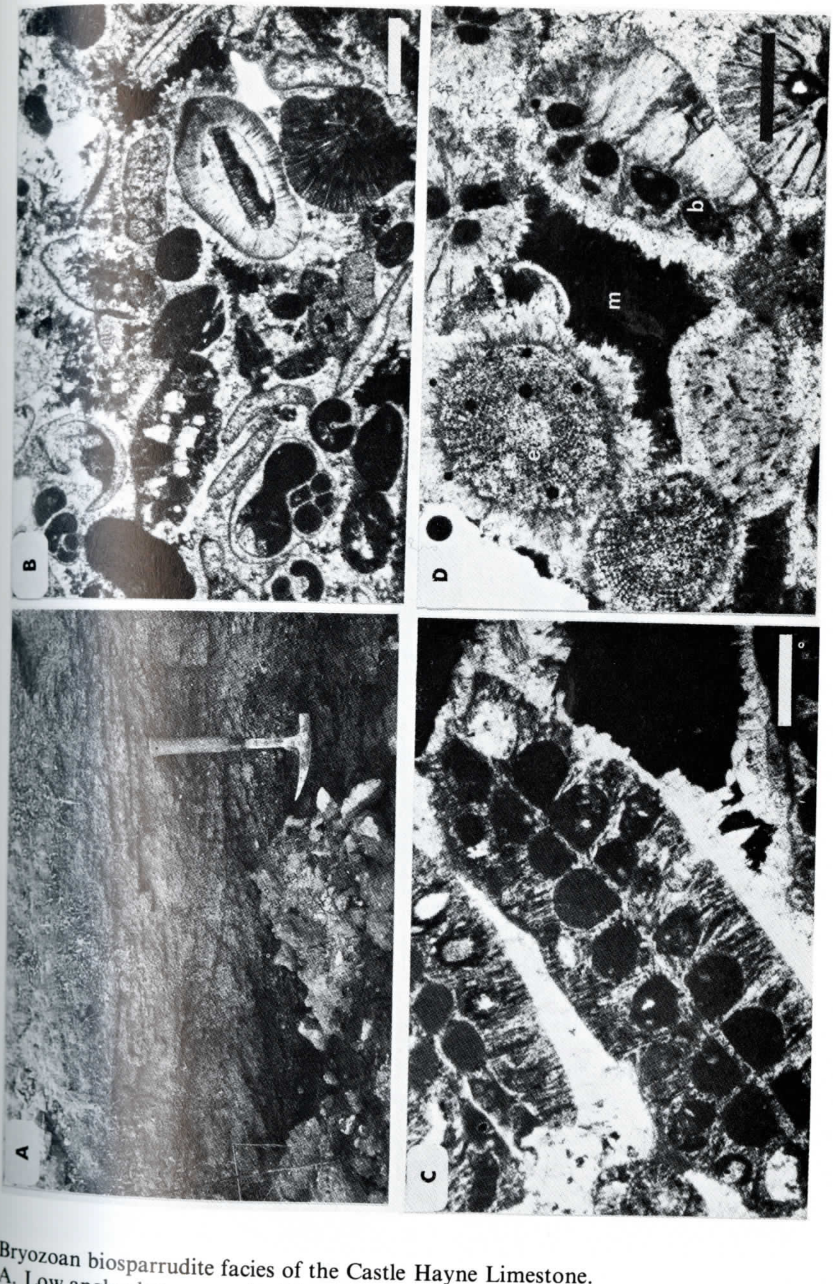
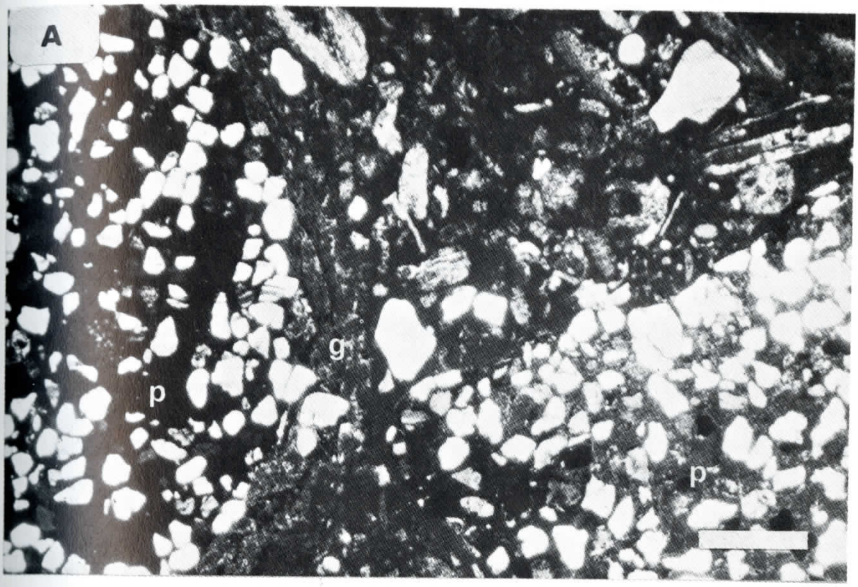


Plate 3. Bryozoan biosparrudite facies of the Castle Hayne Limestone.
A. Low angle planar cross-bedding (hammer is scale).
B. Oolite with allochemical nucleus (x-nichols; scale equals 1.0 mm).
C. Bryozoans cemented by dogtooth spar; note intraparticle micrite in bryozoan zooecia (x-nichols; scale equals 0.5 mm).
D. Bryozoans (b) and echinoids (e) cemented by dogtooth spar; intercrystalline porosity subsequently filled by micrite (m) (uncrossed nichols; scale equals 0.5 mm).

Plate 4. Hollow bryozoan encrustation (left); latex mold of encrustation (center); Recent *Penicillus* (right); note serpulid worms near base (arrow) of latex mold (height of bryozoan encrustation equals 6.8 cm).



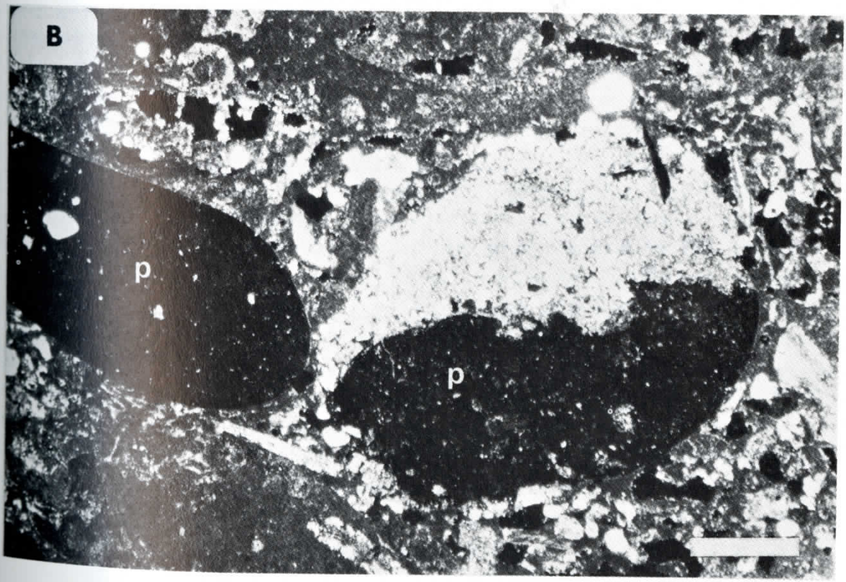


Plate 5. Bryozoan biomicrudite facies of the Castle Hayne Limestone.
A. Bryozoans and pelecypods in a micrite matrix (x-nichols; scale equals 1.0 mm).
B. Bryozoans, foraminifera, an oolite and pelecypods in a micrite matrix (x-nichols; scale equals 1.0 mm).
C. Oolite partially encrusted by a bryozoan; note apparent lack of a nucleus (uncrossed nichols; scale equals 0.5 mm).

pebbles or replacing the micrite or microspar matrix of the pebbles. Generally, only phosphate replaces framework quartz within the pebbles. Degrees of phosphatization range from phosphate laminae which surround the pebbles to phosphate pebbles without remnant fabric (Plate 2).

Although a minor framework component, common allochems include bryozoans, echinoids and pelecypods. Oolites that occur in this facies generally have an allochemical fragment or quartz grain as a nucleus; however, in some cases, there is no apparent nucleus. The oolites generally have distinct, sometimes serrated laminae.

The matrix consists of micrite with very minor amounts of microspar. The terrigenous component includes monocrystalline quartz, detrital phosphate and heavy minerals. The quartz ranges from very coarse to very fine sand size, averaging fine to very fine. Orthochemical constituents consist of glauconite as pellets, laminae, replacement of phosphate and pebble matrices; phosphate as pebbles, phosphatized bone and replacement of terrigenous components and matrices of the pebbles; pyrite as laminae, replacement of pebble matrices, replacement of glauconite pellets and disseminated flakes in the micrite matrix.

Bryozoan Biosparrudite Facies

Distribution and Occurrence. This facies consists of very pale orange (10 YR 8/2), grain supported bryozoans which become medium light gray (N 7) with increasing micrite content. This unit is generally unconsolidated to friable; however, where a diastem separates this facies from the overlying bryozoan biomicrudite facies, the upper portion is dense and well consolidated. The framework is dominanted by bryozoans. Other common elements of the fauna include the sand dollars Protoscutella aff. P. mississippiensis (Twitchell), P. tuomeyi (Twitchell); the cassiduloid Santeelampas oviformis (Conrad); and the brachiopod Terebratulina. The molluscs are a minor component of the fauna; however, the dearth may in part be due to the lack of micrite necessary to form internal molds. Common molluscs include Cubitostrea, Anomia, Chlamys, Gastrochaena and the thin shelled Eburneopecten. Although most of the mega-invertebrates appear to form a life assemblage, the abraded centrodorsals of Microcrinus conoideus Emmons and rounded marginals of Recurvaster indicate that at least some of the biological components have been transported out of the biomicrudite facies. Likewise, many of the bryozoans exhibit abrasion features indicative of transportation. This facies is generally massive; however, low angle planar cross-bedding is a common feature in the downdip portions of this facies (Plate 3a).

In New Hanover County, the bryozoan biosparrudite facies consists of erosional remnants. It is either entirely absent or is separated from the overlying bryozoan biomicrudite facies by a diastem. In Pender, Onslow and Craven counties, this facies interfingers with and

Table 2. Mean Point Count and Insoluble Residue Data for the Bryozoan Biosparrudite Facies of the Castle Hayne Limestone.

Insoluble Residue Analysis by Weight Percent (n=22)

	Mean (%)	Standard Deviation (%)
Carbonate	87.2	8.0
Sand Size	5.6	
Silt + Clay Size	7.2	2.7

Point Count Data (n=20)

Mean Total Rock Porosity	30.8%	
Allochems	(%)	(%)
Bryozoans		26.4
Shell	16.2	
Intraparticle Micrite	7.4	
Intraparticle Spar	0.8	
Intraparticle Porosity	0.5	
Intraparticle Glauconite	0.4	
Echinoids		7.3
Unknown		5.1
Pelecypods		2.9
Shell	2.1	
Moldic Porosity	0.6	
Moldic Spar	0.1	
Gastropods		0.7
Intraparticle Micrite	0.4	
Moldic Porosity	0.3	
Pellets		0.4
Brachiopods		0.3
Foraminifera		0.3
Corals		0.1
Serpulids		T
Ostracods		T
Oolites		T
Scaphopods		T
Terrigenous		
Quartz		3.1
Detrital Phosphate		0.5
Heavy Minerals		T
Matrix		
Micrite		7.8
Porosity		
Interparticle		25.9
Vug		2.3
Channel		1.2
Cement		
Interparticle		14.7
Orthochemical		
Clauconite		1.0
Total		100.0

is gradational with the overlying biomicrudite facies. The maximum thickness observed (12.2 m) occurs in Duplin County.

Generally, this facies becomes sandier toward the axis of the basin (Neuse River) (Brown, 1963) or where disconformable on the Cretaceous Peedee Formation or Paleocene Beaufort Formation. In places this facies is more micritic in the extreme updip portions away from the Carolina fault. This may be a reflection of increased distance from a zone of turbulence causing pockets of micritic sediments within the biosparrudite facies.

The conformable transition between the shallower water bryozoan biosparrudite facies and the deeper water bryozoan biomicrudite facies is delineated by the Carolina fault (Figure 1). Also, the Cape Fear arch appears to have deflected the biosparrudite facies basinward.

Petrography. The major framework components are bryozoans; however, other common allochems include echinoids and pelecypods (Table 2; Plate 3). Oolites (Plate 3b) occurring in this facies are similar to those described from the phosphate pebble biomicrudite facies. They generally have allochemical fragments or sand grains as nuclei and distinct, often serrated laminae. The framework components are incompletely cemented by fine to coarse, generally medium, interparticle dogtooth spar. Occasional fine, intraparticle dogtooth spar is present. Primary depositional micrite increases towards the conformable transition with the overlying bryozoan biomicrudite facies. The terrigenous fraction consists predominantly of well sorted, medium to very fine, averaging fine, monocrystalline quartz sand. Orthochemical constituents consist mostly of glauconite.

When a diastem separates this facies from the overlying biomicrudite facies, at least one generation of filling micrite occurs (Plate 3d). Upchurch (1973) differentiated several generations of filtered, secondary micrite. The presence of first generation interparticle dogtooth spar indicates that this facies was exposed subaerially and was partially cemented in the vadose or phreatic zone. During subsequent transgression, micrite filled the remaining intercrystalline porosity.

Bryozoan Biomicrudite Facies

Distribution and Occurrence. The bryozoan biomicrudite facies generally occurs as an unconsolidated, light gray (N 7) to very pale orange (10 YR 8/2) unit. At exposures where this facies interfingers with the underlying bryozoan biosparrudite facies, it is dense and well lithified. Also, below diastems, this unit is well lithified. Like the underlying bryozoan biosparrudite facies, the fauna is dominated by bryozoans; however, the abundance and diversity of the total fauna increases in this facies (Appendix I).

The delicate, erect escariform bryozoans dominate the fauna; however, cellariform and lunulitiform bryozoans are well represented in this facies. Some of the encrusting bryozoans reveal a tubular

Table 3. Mean Point Count and Insoluble Residue Data for the Bryozoan Biomicrudite Facies of the Castle Hayne Limestone.

Insoluble Residue Analysis by Weight Percent (n=22)	Mean (%)	Standard Deviation (%)
Carbonate	88.1	6.5
Sand Size	4.7	
Silt + Clay Size	7.2	3.4

Point Count Data (n=22)		
Mean Total Rock Porosity	8.9%	
Allochems	(%)	(%)
Bryozoans		21.3
Shell	12.4	
Intraparticle Micrite	8.4	
Intraparticle Porosity	0.4	
Intraparticle Glauconite	0.1	
Intraparticle Spar	T	
Echinoids		7.3
Unknown		2.9
Pelecypods		2.7
Shell	2.0	
Moldic Porosity	0.7	
Foraminifera		0.9
Intraparticle Micrite	0.5	
Shell	0.4	
Brachiopods		0.6
Gastropods		0.3
Intraparticle Micrite	0.1	
Shell	0.1	
Moldic Porosity	T	
Corals		0.1
Ostracods		T
Scaphopods		T
Serpulids		T
Oolites		T
Terrigenous		
Quartz		3.7
Detrital Phosphate		0.1
Heavy Minerals		T
Feldspar		T
Matrix		
Micrite		51.5
Microspar		0.1
Porosity		
Vug		5.8
Channel		2.0
Orthochemical		
Glauconite		0.7
Pyrite		T
	Total	100.0

morphology and hollow centers which indicate that the bryozoans encrusted an erect object that stood above the substrate. A latex mold revealed a morphology strikingly similar to Penicillus (Plate 4). The mold is rounded at one end and gradually flattens at the opposite end. A 2 cm long, broad, shallow groove is apparent along the flattened edges. The mold also reveals serpulid worm tubes which indicates that the bryozoans were not closely appressed to the algae. Voigt (1973) considers this an infallible criterion that the bryozoans encrusted an object, rather than assumed a tubular morphology.

In New Hanover County, the biomicrudite facies contains an abundant sponge fauna, in places, abundant enough to form a densely packed biolithite (Upchurch, 1973; Upchurch and Textoris, 1973).

Although the fauna of the biosparrudite and biomicrudite facies overlap to some degree, many faunal elements appear to be facies controlled. Elements common in the biomicrudite facies generally show signs of transport and subsequent abrasion in the biosparrudite facies (centrodorsals of Microcrinus conoideus Emmons and marginals of Recurvaster). The fauna of the biomicrudite facies does not appear to have been transported; although the pelecypods are generally disarticulated. The presence of massive bedding and the lack of current generated sedimentary structures seems to confirm the lack of strong current activity.

This facies has been the most intensely studied facies of the Castle Hayne Limestone. The type section consists almost entirely of this facies. Probably all of the Castle Hayne Limestone bryozoans studied by Canu and Bassler (1920) were collected from this facies; as well as, most of the molluscs studied by Kellum (1926), the echinoids by Cooke (1959) and the brachiopods by Cooper (1959). Since the bryozoan biosparrudite facies is more easily eroded, the transition from the bryozoan biosparrudite facies to the bryozoan biomicrudite facies approximates the updip limit of the outcropping Castle Hayne Limestone; thus the bryozoan biosparrudite facies is rarely presented in outcrop.

There is some evidence that this facies grades basinward into a sandy, foraminiferal biomicrite; however, subsurface work is necessary to establish its distribution. Therefore, it is not included in the diagnosis of the outcropping Castle Hayne Limestone.

Petrography. Bryozoans are the major framework components (Table 3; Plate 5). Other common allochems include echinoids and pelecypods. The aragonitic allochems now occur as molds.

The terrigenous component consists principally of well sorted, fine to very fine, monocrystalline quartz sand with traces of heavy minerals, feldspar and detrital phosphate. Orthochemical constituents consist of pyrite as disseminated flakes and glauconite. Glauconite also occurs as replacement products of brown bodies in the zooecia of bryozoans.

Dolomite. At the Martin Marietta quarry in New Hanover

County, portions of the bryozoan biomicrudite facies are partially to completely dolomitized. Syndepositional movement along the Cape Fear arch apparently initiated Dorag dolomitization in this area (Harris et al., 1977; Baum et al., 1978a, 1978b).

DEPOSITIONAL ENVIRONMENTS

Paleoecologic interpretations of ancient sedimentary environments cannot be established on any single criterion, nor can an adequate description of the sedimentary history be established without the knowledge of the geometry and lateral facies of the depositional system. Hopefully, a unified study of the petrography, geometry, facies reconstruction and paleoecology has provided an unbiased and accurate interpretation of the depositional environments of the Castle Hayne Limestone.

At the type section of the Castle Hayne Limestone (see Baum et al., 1978, Figure 3), the two major lithofacies of the Castle Hayne Limestone (bryozoan biosparrudite and bryozoan biomicrudite) are separated by a diastem. At this locality, the diastem seems to represent subaerial exposure and/or non-deposition (Plate 3d). The echinoid fauna above and below this diastem also suggests that there is a hiatus (P. M. Kier, per. comm.). Additionally, Upchurch (1973) recognized several diastems within the bryozoan biomicrudite facies exposed at the Ideal Cement Company quarry (locality NH-7 of Baum et al., 1978). Petrographically and faunally, the lithologies above and below these diastems are the same. Basinward (NE), the two lithofacies interfinger and the contact is gradational (localities P-3, ON-3 and CR-5 of Baum et al., 1978). Thus, the Cape Fear arch was apparently active during the deposition of the Castle Hayne Limestone and not only generated local diastems, but initiated Dorag dolomitization (Baum et al., 1978a, 1978b). Basinward (NE), deposition was apparently continuous.

It is assumed that a formation will transgress time (likewise lithofacies). Faunal evidence (V. A. Zullo, per. comm.; P. M. Kier, per. comm.) suggests that the Castle Hayne Limestone does transgress time and evolutionary boundaries; however, formations are defined on lithology. Based on the overall faunal and lithologic similarities, the fundamental depositional system apparently did not change throughout the deposition of the Castle Hayne Limestone; however, periodic tectonic activity affected the basin margins while deposition was continuous away from these tectonic features. The fauna described from each lithofacies may differ with time (evolution); however, the fauna does reflect the ecologic parameters of the environments of deposition at the time of deposition.

Paleoecology

The bryozoan biomicrudite facies contains a more diverse fauna than the updip, shoreward bryozoan biosparrudite facies. Thus, paleoecologic information concerning bathymetry and climate will be interpreted principally from the fauna of the bryozoan biomicrudite facies. Based on stratigraphic position, the bryozoan biosparrudite facies would have been deposited in shallower water than that interpreted for the bryozoan biomicrudite facies.

The abundance of stenohaline forms (bryozoans, brachiopods, corals, echinoids, crinoids) leave no doubt that the Castle Hayne Limestone was deposited in an open environment with normal marine salinity.

The dominance of bryozoans in the Castle Hayne Limestone is unique to the carbonates in the Coastal Plain of North Carolina. Temperature _per se_ does not seem to control the distribution of modern bryozoans (Schopf, 1969), and they are apparently equally distributed in all marine biogeographic zones (Osburn, 1957; Cuffey, 1970); although, some genera appear to be restricted to tropical zones (Cheetham, 1967). Chave (1967) and Milliman _et al._ (1973) alluded to the fact that modern bryozoans appear to be more abundant in non-tropical, generally warm temperate marine biogeographic zones; however, most evidence seems to suggest that bryozoans are more abundant and dominate shelf faunas in areas of low terrigenous sedimentation (Lagaaij and Gautier, 1965; Rucker, 1967; Caulet, 1972; Milliman _et al._, 1973).

The lack of terrigenous sediments in the Castle Hayne Limestone has been noted for years; thus, it is quarried extensively for cement and agricultural lime. The mean soluble carbonate of the bryozoan biosparrudite facies is 87.2% and 88.1% for the bryozoan biomicrudite facies. In areas where the Castle Hayne Limestone contains more terrigenous sediments, the percentage of bryozoans decreases. It would seem that the dominance of bryozoans in the Castle Hayne Limestone was due to low terrigenous influx, rather than other ecologic parameters.

Stach (1936) was the first to attempt to correlate bryozoan zooecial morphology with turbulence (depth). Subsequent work has tended to substantiate his initial proposals (Boillot, 1965; Lagaaij and Gauter, 1965; Ryland, 1967; 1970; Schopf, 1969; Wass _et al._, 1970) which are now frequently used to estimate the bathymetry of ancient sedimentary units (Cheetham, 1963; Askern, 1968; Pedley, 1976).

The bryozoans in the biomicrudite facies are dominated by erect, escariform (includes vinculariform and reteporiform) bryozoans; however, cellariform (_Cellaria_) and lunilitiform (_Lunulites_) are also common zooecial morphologies. The delicate escariform bryozoans are restricted to quiet water, below wave base (Stach, 1936). Schopf (1969) noted a dramatic increase in escariform bryozoans below 35 m. At this depth, effective wave base begins to impinge on the bottom. Cheetham (1963) suggested an optimum bathymetry of 46-183 m for escariform bryozoans.

Lunulitiform bryozoans are unattached, thus they thrive in areas of low turbulence (Stach, 1936; Cheetham, 1963; Ryland, 1967; 1970). The presence of lime mud also suggests a lack of turbulence. Stach suggested an upper limit of lunulitiform bryozoans at 27 m. Cheetham (1963) suggested an optimum bathymetry of 15-107 m.

The articulating cellariform bryozoans are adapted for turbulent zones (Stach, 1936). Cheetham (1963) suggested an optimum bathymetry of 15-46 m.

Caution must be exercised in interpreting depth as a function of bryozoan zooecial morphology. Turbidity, thus depth, is a function of shelf slope (see Irwin, 1965) and prevailing winds. Also, the occurrence of algae and/or marine grasses can distinctly modify these associations (Voigt, 1973). Generally, the maximum diversity of bryozoans occurs at approximately 40 m (Ryland, 1967; 1970; Schopf, 1969) and is apparently enhanced by upwelling of nutrient rich currents at shelf breaks (Wass et al., 1970). The common occurrence of lunulitiform and cellariform bryozoans in the escariform association of the biomicrudite facies of the Castle Hayne Limestone suggests that this facies was deposited at the lower end of the optimum bathymetry for escariform bryozoans. Also, the apparent occurrence of green algae (Penicillus) indicates depths within the photic zone (generally less than 50 m) (Taylor, 1960). Thus, the bryozoan association in the biomicrudite facies of the Castle Hayne Limestone suggests depths of 30-50 m. Cheetham (1963) interpreted a similar Gulf Coast Eocene bryozoan association to have lived at depths of 31-61 m.

Three ahermatypic corals, Balanophyllia, Endopachys and Flabellum, occur in the biomicrudite facies. The concurrence of these three genera suggest depths of 35-600 m, a temperature range of 6.7-26.2°C and low sedimentation rates (Wells, 1956; 1967).

The echinoids are one of the more diverse groups in the Castle Hayne Limestone; however, very little ecologic information can be derived from them. The spatangoids, burrowing forms, are more common in the biomicrudite facies. The sand dollars (Protoscutella) are very common in the extreme updip biosparrudite facies. Heckel (1972) suggested that sand dollars are most common from the intertidal zone to 18 m. The cassiduloids seem to be equally adapted to either facies, although some genera are more common in certain facies. Santeelampas oviformis, like Protoscutella, is essentially restricted to the extreme updip portions of the biosparrudite facies; whereas, Echinolampas appendiculata is essentially restricted to the biomicrudite facies.

The crinoids are associated with the biomicrudite facies. Modern comatulid crinoids, like bryozoans, are most abundant at slope breaks (15-46 m) which are associated with the upwelling of nutrient rich currents (Meyer, 1973; Macurda, 1975). The modern attached crinoid Democrinus has a bathymetric range of 113-2083 m (Clark, 1915; Macurda and Meyer, 1974; 1975). There are no other faunal elements in the Castle Hayne Limestone that suggest such abyssal depths; thus, Democrinus has apparently shifted its optimum depth range since the middle Eocene.

The brachiopods in the Castle Hayne Limestone are essentially facies controlled. Terebratulina is the only genus that appears to be equally adapted to either facies. The pedicle of the modern Terebratulina splits distally into fine rootlets that are able to penetrate both hard substrates, as well as, soft substrates such as algal stems, ascidian tests and horny worm tubes (Rudwick, 1965; Ager, 1967). The ability to adhere to soft substrates probably accounts for its occurrence in fine grained sediments. Argyrotheca, "Terebratula" and Probolarina are essentially restricted to the biomicrudite facies. Argyrotheca can also accommodate soft substrates such as thalluses of algae and colonies of bryozoans (Pajaud, 1974). Modern species of Argyrotheca tend to range from 20-150 m (Davidson, 1877; Pajaud, 1974; Milliman, 1974), although Rudwick (1965) reports them at much greater depths. Argyrotheca johnsoni can reach concentrations of $50/m^2$ at a depth of 60 m (Milliman, 1974).

Although the molluscs are a diversified group in the Castle Hayne Limestone, their identification is inhibited by their occurrence as molds. Except for the pectens and oysters, the remaining molluscs now occur as molds. Extant genera occurring in the Atlantic Ocean have a warm temperate to tropical marine biogeographic zonation and appear to be equally distributed on present day continental shelves (Abbott, 1974). There are several extant genera considered to have only a tropical distribution: Pleurotomaria, Trigonostoma and Agaronia. Modern pleurotomarians are found in tropical waters and generally at a depth not less than 100 m; however, since the Eocene, their depth optimum has shifted from shallow to deep waters (Hickman, 1975). Anomia occurs only in the extreme updip portions of the biosparrudite facies. It is also a common constituent in the basal sediment of the Trent Formation and Belgrade Formation. Its common occurrence at the base of these formations seems to suggest that it is a characteristic shallow water indicator.

The possible occurrence of green algae (Penicillus) suggests depths within the photic zone and a tropical marine biogeographic zonation (Taylor, 1960).

SUMMARY

The estimated bathymetry of the escariform bryozoan association (30-50 m) is substantiated by the distribution of other faunal elements. A maximum depth of 50 m lies within the photic zone, thus within the range of Penicillus. Since the ahermatypic corals have an estimated bathymetry of 35-600 m, it would appear that their occurrence in the Castle Hayne Limestone was at their depth minimum. Other faunal elements, Argyrotheca, molluscs and comatulid crinoids, are abundant within these bathymetric ranges.

The bryozoan biosparrudite facies lies updip and shoreward of the bryozoan biomicrudite facies, thus it was deposited at depths less than 30 m. Supporting data are the abundance of sand dollars (0-18 m). Foraminifera from Natural Well suggest depths of 18-37 m (Copeland, 1964).

The entire fauna represents a warm temperate to tropical marine biogeographic zone; however, several lines of evidence suggest strongly that the Castle Hayne Limestone was deposited in a tropical marine biogeographic zone: the possible occurrence of green algae (Penicillus) (Taylor, 1960); the overall faunal diversity; the great diversity of echinoids (Durham, 1966). Foraminifera also suggest a warmer climate than presently at this latitude (Copeland, 1964). Thus, the lime mud less than 4μ was probably generated by large standing crops of calcareous green algae.

The transition between the blanket carbonate sands of the biosparrudite facies and the deeper water bryozoan lime muds of the biomicrudite facies is essentially delineated by the Carolina fault. This transition suggests that the Carolina fault created a shelf break that initiated and directed currents onto the platform (see Ball, 1967). The slope break also initiated the upwelling of nutrient rich waters. Seaward of the fault, lower energy conditions and nutrient rich waters provided an excellent habitat for a diverse population of sessile, filter feeding organisms such as bryozoans, brachiopods and crinoids. Leeward of the fault, currents transported, sorted and concentrated allochems on the platform. Some elements in the biosparrudite facies exhibit abrasion due to transportation out of the micrite environment; however, the higher energy conditions on the platform provided a suitable habitat for many organisms (Anomia, Protoscutella and Eburneopecten). The lack of sedimentary structures in the extreme updip portion of the biosparrudite facies is probably a reflection of sediment mixing by burrowing organisms (see Ball, 1967).

Although the Neuse fault has been periodically active from the Lower Cretaceous to the Pleistocene (Harris et al., 1979), the presence of the Neuse fault is not reflected in the lithofacies distribution of the Castle Hayne Limestone. Apparently, the Neuse fault was not tectonically active, nor a positive feature during the deposition of the Castle Hayne Limestone.

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APPENDIX I

Fauna From The Bryozoan Biomicrudite Facies of The Castle Hayne Limestone (see Canu and Bassler, 1920, for list of bryozoans).

GASTROPODS

Agaronia
Architectonica
Athleta
Calyptraea
Caricella
Conus
? Crucibulum
Cypraea
Diodora
Emarginula
Epitonium
Ficopsis
Fusimitra
Galeodea
Marginella
? Neosimnia
Phalium
Pleurotomaria
? Pseudoliva
Puncturella
? Sconsia (Doliocassis)
Serpulorbis
? Strombus
Tenagodus cf. T. vitis (Conrad) 1833
Trigonostoma
Turritella
Xenophora

CEPHALOPODS

Aturia alabamanensis (Morton) 1834
"Belosaepia"
Eutrephoceras carolinensis Kellum 1926

PELECYPODS

Aracea
Barbatia cf. *B.* (*Cucullaearca*) *cuculloides* (Conrad)
Cardiinae
Chama
Chlamys deshayesii (Lea) 1833
Chlamys membranosus (Morton) 1834
Corbula
Crassatella aff. *C. alta* Conrad 1832
Crassatellinae
Cubitostrea cf. *C. sellaeformis* Conrad 1832
? *Glossus*
Gryphaeostrea cf. *G. subeversa* Conrad 1866
? *Lirodiscus*
Modiolinae
Ostrea falco Dall 1895
Ostrea trigonalis Conrad 1854
Pecchiolia dalliana Harris 1919
Pholadomya
Plicatula filamentosa Conrad 1833
Solena
Spondylus lamellacea Kellum 1926
? *Venericardia*

SCAPHOPODS

Dentalium

BRACHIOPODS

Argyrotheca
Probolarina holmesii (Dall) 1903
P. salpinx (Dall) 1903
"*Terebratula*" *wilmingtonensis* Lyell and Sowerby 1845
Terebratulina

CORALS

Balanophyllia
Endopachus
Flabellum

ECHINOIDS

Agassizia (*Anisaster*) *wilmingtonica* Cooke 1942
Arbacia
Cidaris pratti (Clark) 1915
Echinocyamus parvus (Emmons) 1858
Echinolampus appendiculata Emmons 1858
Eupatagus (*Gymnopatagus*) *carolinensis* (Clark) 1915
Eurhodia rugosa (Ravenel) 1848
Linthia hanoverensis Kellum 1926
L. wilmingtonensis Clark 1915

ECHINOIDS
(continued)

Maretia subrostrata (Clark) 1915
Periarchus lyelli (Conrad) 1834
Phymosoma cf. P. dixie Cooke, 1941
Protoscutella conradi (Cotteau) 1891
Rhyncholampas cf. R. conradi (Conrad) 1850

OPHIUROIDS

Euryale (cf. Asteronyx)
Ophuirinae- broad and narrow arm species

ASTEROIDS

?Metopaster
Recurvaster

CRINOIDS

Amphorometra
Democrinus
Microcrinus conoideus Emmons 1858

PRELIMINARY INVESTIGATION OF THE PALEOMAGNETISM OF FLORIDA CENOZOIC CARBONATES

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ABSTRACT

During 1978-1979 a total of 157 limestone cores were drilled for paleomagnetic analyses from 22 sites that span a wide geographic and temporal (late Eocene to Recent) range in Florida. Laboratory measurements of these specimens (cores) using a cryogenic magnetometer revealed natural remanent magnetization (NRM) intensities ranging on the order of 10^{-6} emu/g to 10^{-9} emu/g. The specimens were demagnetized using alternating field (a. f., 30 oe, 85 oe, 150 oe, 300 oe) and thermal (220° C, 300° C, 350° C) techniques. There was less than an order of magnitude decrease in specific intensity after demagnetization. In many cases these levels of demagnetization satisfactorily removed viscous components of magnetization (VRM). Some other sites will require demagnetization at higher fields and temperatures. At least two sites demonstrate chemical remanent magnetization (CRM) as a result of the presence of hematite. Sites of normal and reversed polarity are identified in this study. There are no *a priori* criteria that can be used to predict if either a. f. or thermal demagnetization will be more effective in removing secondary components of magnetization such as VRM. This preliminary survey demonstrates the potential of applying paleomagnetic techniques to an analysis of earth history as recorded in Florida carbonates.

INTRODUCTION

During the last three decades the study of paleomagnetism of sediments has had numerous applications in the earth sciences. The early work on paleomagnetism of sediments was limited in scope because only the more strongly magnetized sediments (mostly "red beds") could be analyzed with the laboratory equipment available at that time (Cox and Doell, 1960). More recently, there have been paleomagnetic studies of a variety of sediments because of the increased sensitivity of the laboratory instrumentation. Of particular relevance to the present report was the advent of the crogenic magnetometer (Goree and Fuller, 1976) which allows rapid analyses of very weakly magnetized sediments such as limestones. In recent years there has been increased interest in the paleomagnetism of Phanerozoic limestones to solve a diversity of geological problems (e.g., McElhinny and Opdyke, 1973, Shive and Frerichs, 1974, Martin, 1975, Lowrie and Alvarez, 1975, Gose and Swartz, 1977).

To date, the only published paleomagnetic results from Florida involve an analysis of Recent unconsolidated lime muds from the Florida Keys (Jowett and Pearce, 1977). When these muds were dried, they exhibited an NRM (natural remanent magnetization) consistent with the present field and intensities on the order of 10^{-7} emu/cm^3.

The purpose of this report is to demonstrate the potential application of paleomagnetism to Cenozoic carbonates of Florida. This pilot study will then serve as a basis for focusing on problems such as stratigraphic and lithofacies correlations, paleomagnetic pole positions, and evaluation of tectonic stability of the Florida peninsula. The present study represents the only published results of paleomagnetism of pre-Recent Florida rocks.

Acknowledgments

This research was supported by Seed Money Grants from the University of Florida Division of Sponsored Research. We are most grateful to Dr. Wulf Gose for availing the University of Texas Marine Science Institute Paleomagnetic Laboratory facilities to us and for invaluable advice. Jon A. Baskin and Graig D. Shaak critically read the manuscript and offered helpful suggestions for its improvement.

GEOLOGIC SETTING

The Florida plateau, including the emergent peninsula and the west Florida continental shelf, consists of a thick wedge of late Mesozoic and Cenozoic sedimentary rocks. Only Cenozoic rocks, however, are exposed, and their surficial nature and distribution have been described by Puri and Vernon (1964) and Stringfield (1966). Carbonate

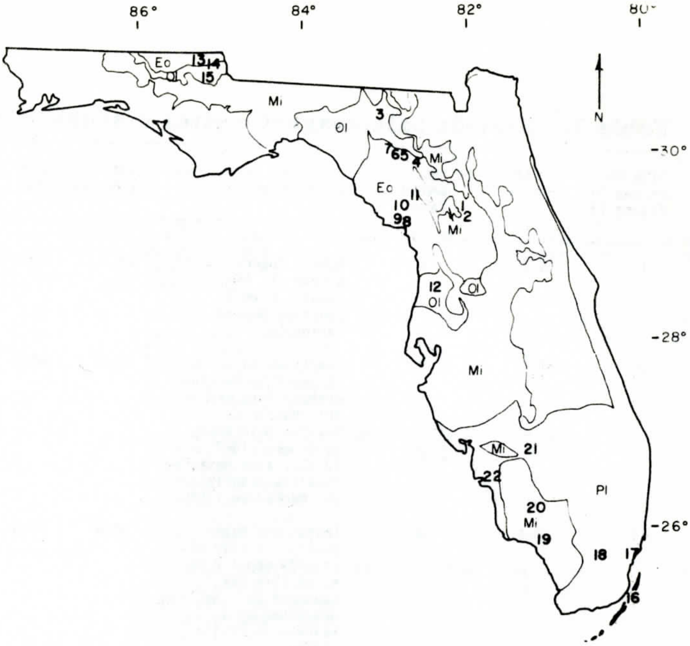

Figure 1. Florida paleomagnetic sites discussed in this report. Site numbers correspond to descriptions in Table 1. Key: Eo, Eocene; Ol, Oligocene; Mi, Miocene; Pl, Pleistocene; other areas, Recent and Pleistocene. After Puri and Vernon (1964).

rocks forming the surface of the peninsula range (Figure 1) in age from Eocene (Claiborne Stage) to Pleistocene (Puri and Vernon, 1964).

Exposures in the southern one-third of the peninsula consist of the Tamiami Formation of Miocene age (Choctawhatchee Stage) and the Anastasia Formation, Miami Oolite, Key Largo Limestone, Fort Thompson Formation, and Caloosahatchee Formation of Pleistocene age.

In the central part of the peninsula the Ocala Group of limestones (Jackson Stage) and Avon Park Limestone (Claiborne Stage) are of Eocene age. To the north are outcrops of the Suwannee and Marianna (in the panhandle) limestones, both of Oligocene age. Miocene units include the phosphoritic Hawthorn and Bone Valley Formations (Alum Bluff Stage) and the St. Marks Formation (Tampa Stage).

Although topographic irregularities in erosion surfaces may have caused disconformities between certain units, the various surficial formations are essentially devoid of tilting or structural deformation. Some post-Oligocene orogenic uplift has been proposed (e.g. Hoyt, 1969) as has tilting during the Eocene (Winston, 1976). Shoreline transgressions and regressions and the general emergence of stratigraphic sections during the late Cenozoic, however, can be attributed to eustatic sea-level changes (e.g. Peck et al., 1979).

Table 1. Florida paleomagnetic sites discussed in this paper.

Site No. (on map in Figure 1)	Site Abbrev.	No. Samples Taken	Site Description	Lat (°N)	Long (°W)	Unit	Age
1	N-1	7	Norris Quarry on W side rt. 441, about 1.5 km S junction 301-441 Marion Co.	29.3	82.8	Ocala Grp., Crystal River Fm.	late Eocene
2	MA-1	7	Quarry on SW corner of junction Martin-Anthony Road and rt. 301, Marion Co. Section described by Brooks (1967, p. 22-23), also detailed locality description in MacFadden (1980).	29.3	82.2	Ocala Grp., Crystal River Fm.	late Eocene
3	LO-1	6	Shands and Baker Quarry. E side of rt. 129 about 2 km N. of Live Oak, Suwannee Co. Section described by J. S. Waldrop in Frailey (1978).	30.3	83.0	Suwannee Limestone	late Oligocene-early Miocene
4	HS-1	7	Outcrop along N side rt.27 about 6 Km SE Fort White, Columbia Co.	29.7	82.6	Ocala Grp., Crystal River Fm.	late Eocene
5	BR-1	7	Abandoned limestone pit about 8 km E of Branford, Suwannee Co.	29.9	82.9	Ocala Grp., Crystal River Fm.	late Eocene
6	BR-2	7	NE corner Hatch Quarry, about 3 km E Branford on rt. 27, Suwannee Co. Dolomitic	29.9	82.9	Ocala Grp., Crystal River Fm.	late Eocene
7	BR-3	7	Abandoned Quarry on W side rt. 129 about midway between Branford and O'brien, Suwannee Co.	30.1	82.9	Ocala Grp., Crystal River Fm.	late Eocene
8	IN-1	7	"Bed 3" in limerock pit near type section of Inglis Fm., Levy Co. Stop 2 in SEGS Guidebook No. 18 (1976).	29.1	82.7	Ocala Grp., Inglis Fm.	late Eocene
9	IN-2	7	Type section Inglis Fm., Inglis Boat Basin on N bank Withlacoochee River just S of rt. 40, Levy Co. Stop 3 in Puri and Vernon (1964).	29.1	82.7	Ocala Grp., Inglis Fm.	late Eocene
10	GH-1	7	Dixie Lime Quarry, E side rt. 19 and S of Gulf Hammock, Levy Co. Dolomitic, Stop 3 in SEGS Guidebook No. 18 (1976).	29.3	82.7	Avon Park Ls.	Eocene
11	W-1	7	Pit on E side rt. 335A, 1 km N of Williston, Levy Co.	29.4	82.4	Ocala Grp., Williston Fm.	late Eocene
12	B-1	7	Florida Crushed Stone Quarries, Leesville, Hernando Co.	28.5	82.3	Suwannee Limestone	late Oligocene-early Miocene

Table 1. continued.

Site No. (on map in Figure 1)	Site Abbrev.	No. Samples Taken	Site Description	Lat. (°N)	Long (°W)	Unit	Age
13	J-1	7	Roadcut E side rt. 73 about 6 km N of junction rt. 90, Jackson Co.	30.8	85.3	? Ocala Grp., Crystal River Fm.	late Eocene
14	J-2	7	Abandoned Limerock Quarry near Springfield Church, Jackson Co. Stop 14 in Puri and Vernon (1964).	30.8	85.3	Ocala Grp., Crystal River Fm.	late Ecoene
15	J-3	5	Type section of Marianna Ls. on N side rt. 90 in Marianna, Jackson Co. See Stops 25 and 26 in Puri and Vernon (1964).	30.7	85.3	Marianna Limestone	[illegible]gocene
16	KY-1	7	Key Largo, about 3 km S of Pennecamp State Park, coral reef facies exposed in canal, Monroe Co.	25.2	80.4	Key Largo Limestone	Pleistocene
17	SM-1	8	Banks of Coral Gables Waterway in South Miami about 3 km E of rt. 1 at junction LeJeune and Sunset Roads, Dade Co. Stop 84 in Puri and Vernon (1964).	25.7	80.4	Miami Oolite	Pleistocene
18	TT-1	8	Banks of canal just N of Tamiami Trail (rt.41) at E entrance Big Cypress Nature Preserve, Dade Co. Dolomitic.	25.8	80.8	Miami Oolite	Pleistocene
19	TT-2	6	Exposures on S side Tamiami Trail about 5 km E of Carnestown, Collier Co.	25.8	81.4	Tamiami Limestone	Miocene
20	SL-1	7	Roadcrop on rt. [illegible] about 100 m [illegible] of rt. 29 and about [illegible] km N of [illegible]	[illegible]	[illegible]	[illegible]	[illegible]
21	[illegible]	[illegible]	Banks [illegible] Ortona [illegible] about [illegible] 2 km [illegible] Glades [illegible] in Puri and Vernon [illegible] 1964	[illegible]	[illegible]	[illegible]	[illegible]
22	FM-[illegible]	8	Roadcrop on [illegible] side rt. 82 about [illegible] km E of Fort Myers, Lee Co. ? Dolom[illegible] Stop 83 in Puri and Vernon 1964	[illegible]	[illegible]	[illegible]	[illegible]

PROCEDURES

During 1978 and 1979, a total of 157 short (50 mm or less) cores were drilled at 22 sites using a commercially available gasoline-powered portable rock drill. At each site five to eight separately oriented cores were taken at the surface within a lateral extent of less than 5 m and with a vertical separation of less than 2 m. The cores were randomly oriented in various directions to provide a check on internal consistency of the laboratory data. Because this research is a preliminary survey, sites were drilled in a variety of Florida carbonates (mostly limestones) that span a wide time (late Eocene-Pleistocene) and geographic range (Figure 1, Table 1).

In the laboratory external weathering rinds were removed and these specimens were cut and sanded into standard 25.4 mm long cylinders. In order to accurately calculate specific intensities for individual suites of specimens with variable porosities, all specimens were weighed. Typical densities were approximately 2.0 gm/cm^3, and values for intensities of magnetization in emu/cm^3 are simply the product of densities and specific intensities (in emu/gm). All paleomagnetic measurements and demagnetizations were done by the authors at the University of Texas Marine Science Institute at Galveston. The equipment used at this laboratory included a Superconducting Technology cryogenic magnetometer, a Tektronix in-line mini-computer (for rapid data analysis and storage on magnetic tapes), an alternating field (a. f.) demagnetizer, and a thermal demagnetizer. All rock specimens were first measured for their NRM, demagnetized one or more times (either using a. f. or thermal techniques, or in a few cases, both), and their magnetization was remeasured after each demagnetization. Ten successive measurements were taken in each of three axes for NRM determinations and after each demagnetization using the cryogenic magnetometer. These data were then analyzed using the in-line mini-computer and a resultant vector was obtained.

The laboratory analyses were designed to characterize the magnetic properties and behavior of a wide range of Florida carbonates. This survey was undertaken to concentrate on those localities and lithologies that demonstrate "favorable" paleomagnetic characteristics for future studies (see conclusions).

MAGNETIC MEASUREMENTS

NRM. All specimens were measured for their NRM. The site mean specific intensity (order of magnitude) for NRM characteristically ranges from 10^{-7} emu/g to 10^{-9} emu/g (Table 2). Two sites (J-2 and KY-1) were deliberately collected because of their obvious hematitic staining. These two sites demonstrate NRM specific intensities in the 10^{-6} emu/g range. It was found that consistency of NRM (5 to 8 specimens) for sites in the 10^{-6} emu/g to 10^{-8} emu/g range was within an order of magnitude. The consistency of NRM measurements for the 4 sites in the 10^{-9} emu/g range (N-1, MA-1, LO-1, BR-2) was more variable (approximately two orders of magnitude).

It is assumed that the primary NRM of most sites results from

Table 2. Mean Intensity Range (order of magnitude in emu/g) for NRM and Successive demagnetization Treatments for the 22 Sites Discussed in this Report. See Table 1 for Site Descriptions.

Site	NRM	Demagnetization						
		30 oe	85 oe	150 oe	300 oe	220^{0} C	300^{0} C	350^{0} C
N-1	10^{-9}	–	–	–	–	10^{-9}	–	–
MA-1	10^{-9}	–	–	–	–	10^{-9}	–	–
LO-1	10^{-9}	–	–	–	–	10^{-9}	–	–
HS-1	10^{-8}	–	–	–	–	10^{-9}	–	–
BR-1	10^{-8}	–	–	–	–	10^{-9}	–	–
BR-2	10^{-9}	–	–	–	–	10^{-9}	–	–
BR-3	10^{-8}	–	–	–	–	10^{-9}	–	–
IN-1	10^{-8}	–	–	–	–	10^{-8}	–	10^{-8}
IN-2	10^{-7}	10^{-7}	–	–	–	–	–	–
GH-1	10^{-7}	–	–	–	–	10^{-8}	10^{-8}	–
W-1	10^{-7}	–	–	10^{-8}	–	–	–	–
B-1	10^{-8}	–	–	–	–	10^{-8}	–	–
J-1	10^{-7}	–	10^{-8}	–	–	–	–	–
J-2	10^{-6}	–	–	–	–	10^{-6}	–	–
J-3	10^{-8}	10^{-8}	–	–	10^{-8}	–	–	–
KY-1	10^{-6}	–	–	–	–	10^{-7}	10^{-7}	–
SM-1	10^{-7}	–	10^{-8}	–	–	–	–	–
TT-1	10^{-8}	–	–	–	–	10^{-8}	–	–
TT-2	10^{-8}	–	–	10^{-8}	–	–	–	–
SL-1	10^{-7}	–	–	10^{-8}	–	10^{-7}	–	–
OL-1	10^{-7}	10^{-7}	–	–	10^{-8}	–	–	–
FM-1	10^{-7}	–	–	–	–	10^{-8}	–	–

DRM (detrital remanent magnetization) or pDRM (post-depositional detrital remanent magnetization, sensu Verosub, 1977). It is clear that there are at least two secondary components of NRM recorded in these limestones. All sites demonstrate a secondary VRM (viscous remanent magnetization) apparently caused by the present field. The stereographic projections of NRM for sites J-3 (Figure 2a) and MA-1 (Figure 3a) show VRM overprinting for what is interpreted to be sites

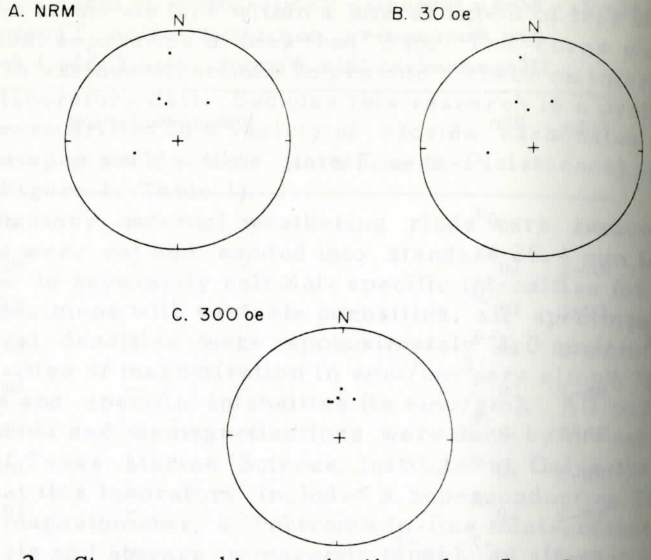

Figure 2. Stereographic projections for 5 paleomagnetic specimens from site J-3, Oligocene Marianna Limestone, Jackson County. a. NRM b. A. f. demagnetized at 30 oe. c. A. f. demagnetized at 300 oe. Crosses represent positive inclinations.

of normal and reversed polarity, respectively. At least two sites (J-2, Figure 4; and KY-1) demonstrate a CRM (chemical remanent magnetization) apparently caused by the presence of hematite. At this time, it is not possible to determine the time of acquisition of CRM.

A. F. Demagnetization. All specimens from 8 sites were subjected to a. f. demagnetizations of various combinations of 30 oe, 85 oe, 150 oe, or 300 oe (Table 2). At 30 oe there was neither a significant drop in intensity nor clustering of the individual specimen vectors (Figure 2b). At higher a. f. demagnetizations (85 oe to 300 oe) there was about an order of magnitude or less drop in specific intensity (Table 2) and, as exemplified by site J-3 (Figure 2c), a statistically significant clustering of magnetic vectors for all specimens (N=5, R=5, k=103.5, α_{95}=7.6°).

Thermal Demagnetization. All specimens from 15 sites were subjected to thermal demagnetization at 220° C (Table 2). In addition, 3 sites (IN-1, GH-1, and KY-1) were subjected to thermal demagnetizations of either 300° C or 350° C. There was a significant decrease in specific intensity after demagnetization at 220° C relative to the NRM. In all cases this drop was less than an order of magnitude. As exemplified by site MA-1 (Figure 3), thermal demagnetization removed part of what appears to be a present-field normal VRM overprint and the

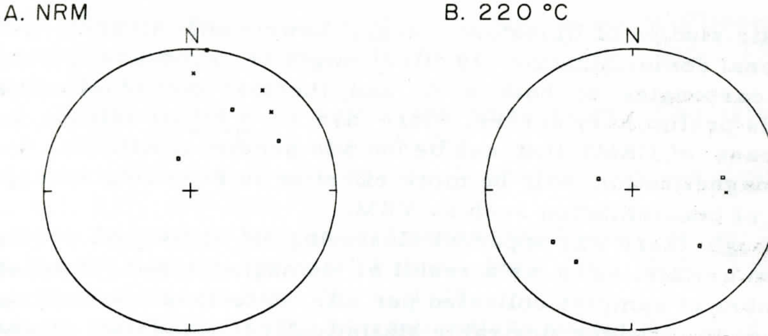

Figure 3. Stereographic projections for 7 paleomagnetic specimens from site MA-1, late Eocene Crystal River Formation, Marion County. a. NRM b. Thermally demagnetized at 220° C. Crosses represent positive inclinations and boxes represent negative inclinations.

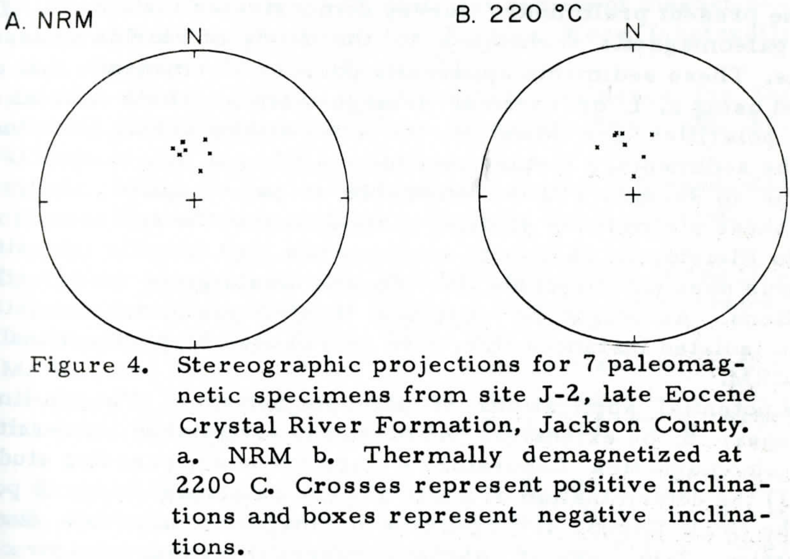

Figure 4. Stereographic projections for 7 paleomagnetic specimens from site J-2, late Eocene Crystal River Formation, Jackson County. a. NRM b. Thermally demagnetized at 220° C. Crosses represent positive inclinations and boxes represent negative inclinations.

individual specimen vectors moved to the southern hemisphere, indicating a reversed polarity for the site.

Discussion. In certain cases, either a. f. or thermal demagnetization techniques appear useful for removing viscous components of magnetization. Many of the specimens responded to the demagnetizations performed in this study. However, some other specimens in this study will probably require a. f. demagnetization at higher fields or thermal demagnetization at higher temperatures in order to satisfactorily remove parts of the VRM. In addition, as has been done in other paleomagnetic studies of limestones (e. g., Lowrie and Alvarez, 1975, Gose, personal communication, 1979), it might be necessary to subject the Florida carbonates to both a. f. and thermal demagnetization. Based on this preliminary survey, there are no a priori criteria (except in the case of CRM) that can be used to predict if either a. f. or thermal demagnetization will be more effective in removing secondary components of magnetization such as VRM.

Although there was apparent clustering of individual specimen vectors from certain sites as a result of demagnetization, it appears that the number of samples collected per site (5 to 8) is not sufficient in some cases to produce desirable statistically filtered data (Fisher, 1953). As Gose and Swartz (1977) have demonstrated, perhaps as many as 25 samples per site are necessary if data such as paleomagnetic pole positions are desired.

CONCLUSIONS

The present preliminary survey demonstrates the feasibility of applying paleomagnetic techniques to the study of Florida Cenozoic limestones. These sediments apparently possess a remanence that can be isolated using a. f. or thermal demagnetization. Both normal and reversed polarities are found in the sites drilled during this study. Much of the sedimentary history recorded in Florida limestones, i. e., late Eocene to Recent, will be amenable to paleomagnetic analysis. Based on these preliminary studies, it appears that the southern Florida Miocene to Pleistocene carbonate sequence has high specific intensities although this does not preclude the Eocene Ocala group from further investigations. As might be expected, post-depositional hematitic staining in isolated instances (i. e. site J-2) causes high intensities because of CRM.

The potential applications of paleomagnetism of Florida limestones appear to be extensive. At the newly established University of Florida Paleomagnetics Laboratory we are presently pursuing studies such as: 1) the determination of a late Eocene paleomagnetic pole position for the Ocala Group, 2) diagenesis of magnetic minerals during dolomitization, 3) the use of magnetic reversal stratigraphy for correlation between the marine and non-marine intergradations in the late Cenozoic of Florida.

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PETROLOGY OF THE RICH MOUNTAIN ULTRAMAFIC BODIES AND ASSOCIATED ROCKS, WATAUGA COUNTY, NORTH CAROLINA

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ABSTRACT

Ultramafic rocks consisting of dunite, harzburgite, and orthopyroxenite are found in two small masses surrounded by amphibolite of the Ashe Formation on Rich Mountain in the Blue Ridge Province of North Carolina. Dunite, the dominant rock type, is gradational with harzburgite. Discrete banks of orthopyroxenite, 1 to 30 cm thick, are found in one of the dunite masses. Mineral assemblages in the amphibolites (hornblende + plagioclase + quartz + sphene $\pm$ epidote $\pm$ garnet $\pm$ clinopyroxene $\pm$ opaque) and the ultramafic (olivine + tremolite + orthopyroxene and olivine + tremolite + talc) typify the middle amphibolite facies of regional metamorphism for these bulk compositions. The ultramafic rock contains significant amounts of magnesite, together with some mineral assemblages such as olivine + orthopyroxene + anthophyllite, olivine + orthopyroxene + talc, and orthopyroxene + talc + magnesite, indicating a significant CO_2 component in the fluid that accompanied metamorphism of the ultramafic rocks. This high CO_2 content may have resulted in the stabilization of orthopyroxene at lower temperatures than anthophyllite in the system $CaO\text{-}MgO\text{-}SiO_2\text{-}H_2O\text{-}CO_2$.

Textures in the olivine-rich rocks indicate at least two and possibly three periods of deformation have resulted in recrystallization of the ultramafic rock. A penetrative deformation resulting in a microscopic foliation has recrystallized some dunite. Static recrystallization without the influence of a deforming force has resulted in a polygonal olivine fabric with 120° grain boundaries. None of the olivine grains show kink bands or undulatory extinction. Porphyroblasts of orthopyroxene with kink bands and showing undulatory extinction are found in the harzburgite and orthopyroxenite. These deformed orthopyroxene grains may represent a third deformation or recrystallization during one of the events that produced the olivine fabrics. Emplacement of

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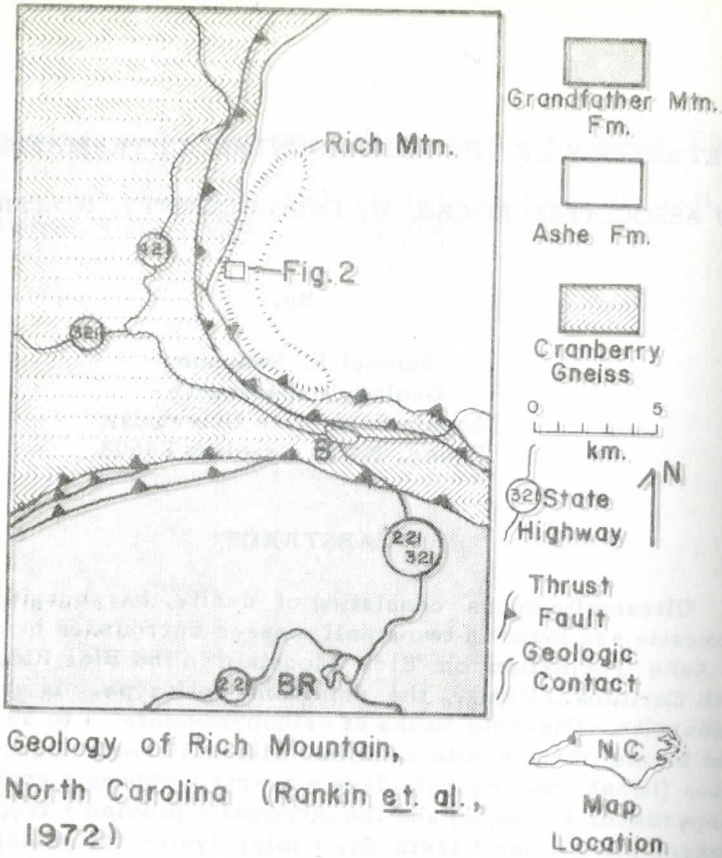

Figure 1. Geologic setting of Rich Mountain. B is Boone and BR is Blowing Rock.

the ultramafic rocks predates the major Ordovician regional metamorphism in the Blue Ridge of North Carolina.

INTRODUCTION

Ultramafic rocks found near the summit of Rich Mountain in north-central Watauga County (Figure 1) are typical representatives of a belt of ultramafic rocks found in the Blue Ridge belt of western North Carolina (Larabee, 1966). At least two distinct masses of ultramafic rock are found on Rich Mountain (Figure 2) and each of these masses is represented by a slight topographic high. The northern body of ultramafic rocks (body 1) is very poorly exposed as it underlies a pasture. Piles of rocks from the pasture have been heaped on the ultramafic outcrops, apparently when the pasture was cultivated. Removal of loose rock from the pasture made mapping of the boundaries between the ultramafic body and the country rocks difficult. Exposures on the southern body of ultramafic rock (body 2) are much better, but the contacts with the country rock are not exposed. Dunite is the dominant rock type in both bodies (Hearn et al., 1977a). Small, discontinuous layers of orthopyroxenite are found in dunite body 2.

Pratt and Lewis (1905) first mapped the ultramafic rocks on Rich Mountain and they also noted the occurrence of enstatite pyroxenite in the ultramafic bodies. In an effort to better outline dunite body 1, a geochemical survey for Ni was made from soil samples collected in the pasture (Hearn et al., 1977b; Callahan et al., 1978). In addition the same authors report on the results of magnetic and radiometric surveys over Rich Mountain dunite body 1. Results of these surveys suggest that the dunite body in the pasture may consist of two distinct masses. However, this suggestion could not be confirmed by geologic mapping (Hearn et al., 1977a) owing to the limited exposure.

Results of a preliminary examination of the petrography of the Rich Mountain dunites and associated country rocks have been published previously (Hearn et al., 1977a). The purpose of this paper is to expand on the relation of mineral assemblages in the dunite and the country rocks. In addition, since publication of the initial petrographic report, bands of pyroxenite were discovered in dunite body 2 and this study will report on the relation of mineral assemblages in the pyroxenite and the dunite. The presence of two distinct ultramafic bulk compositions should give a more complete picture of the petrogenesis of the ultramafic rocks.

Acknowledgments

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METHODS OF INVESTIGATION

Geologic mapping of the Rich Mountain ultramafic rock bodies and surrounding country rocks was done during the geochemical and geophysical investigations of body 1. Because of the limited number of rock outcrops, no attempt was made to map structures within the ultramafic bodies. Samples of ultramafic rock were collected during the detailed investigation of body 1 (Hearn et al., 1977b; Callahan et al., 1978) and during a reconnaissance geophysical study of body 2. Many of the samples taken from body 1 were collected from the debris piles in the pasture so that a representative sample of the ultramafic body could be obtained. Samples from body 2 were all collected from outcrops.

One problem in mapping country rock was to determine which outcrops were in place and which rock masses had been subjected to downslope movement. In the Blue Ridge belt, mass wasting, facilitiated

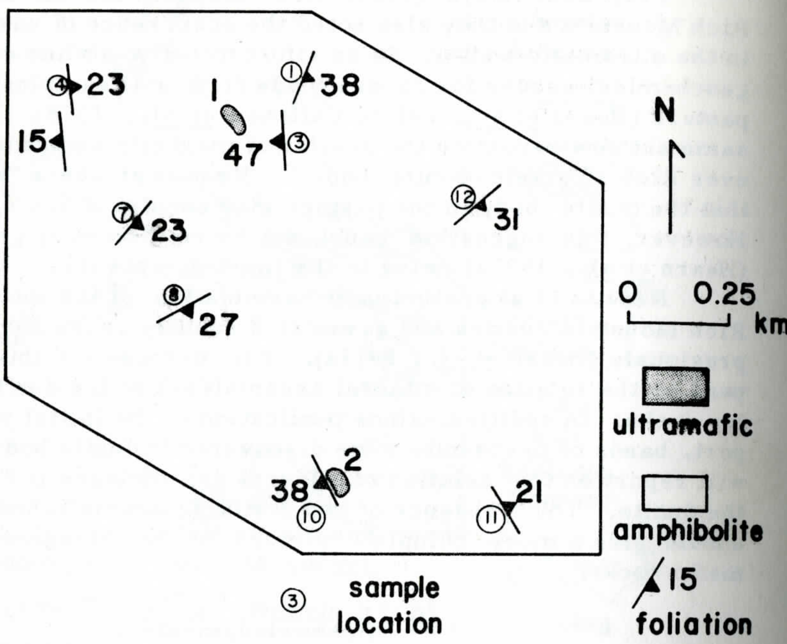

Figure 2. Geologic map of the area surrounding the Rich Mountain ultramafic bodies (numbered 1 and 2 in text).

by steep slopes, heavy rainfall, and frequent freeze-thaw cycles, is very common. In the Rich Mountain area the entire regolith is often found to be moving downhill carrying blocks of country rock that are tens of meters across. This mass wasting creates problems in interpretation of foliations in those terrains with limited outcrops. In an attempt to assure that foliations depicted on Figure 2 represent measurements on bedrock outcrops and not blocks in the regolith, several attitudes were determined within a 50 meter radius and compared to the initial measurement. Where the results of the several measurements were found to yield widely different values, the "outcrop" was considered to be a block in the regolith and the foliation was not plotted.

Foliations shown on Figure 2 represent consistent results on several measurements within 50 meters of the outcrop. Despite these precautions, foliations shown on Figure 2 do not show any consistent pattern and this may result, at least in part, from measurements made on blocks in the regolith.

Modal analysis was done on thin sections of the ultramafic and country rocks. When a foliation was present in the samples, the thin sections were cut perpendicular to this planar structure. Point counting was done with a spacing between points of 1.0 mm. This spacing is larger than the average grain size of the dunites and is about the same magnitude as the grain size in most of the amphibolites. Use of this point spacing on a standard thin section resulted on modal analysis based on 400-700 points.

COUNTRY ROCKS

Metamorphic Rocks

Plagioclase hornblende gneiss (amphibolite) is the dominant rock type surrounding the ultramafic bodies (Figure 2). Within the gneiss are a few layers which are more quartzo-feldspathic and range in thickness from 1 to 30 cm. These layers are discontinuous along strike, even at the outcrop scale. Layering within the gneiss is probably related to differentiation of mafic and felsic components during metamorphism rather than any relic bedding.

Mineral assemblages in the country rocks are indicative of the middle amphibolite facies of the Barrovian Facies Series of regional metamorphism. In the amphibolites, the dominant assemblage is hornblende + plagioclase (An_{24}-An_{34}) + quartz + sphene $\pm$ epidote $\pm$ clinopyroxene $\pm$ garnet $\pm$ opaque (Table 1). Some zoisite, found in small shear zones within the amphibolite, perhaps indicates recrystallization during a period of retrograde metamorphism. Clinopyroxene does not appear to be stable in the amphibolites, as evidenced by overgrowths of hornblende and a turbid appearance in the pyroxene. All other phases appear to be in textural equilibrium. Assemblages in pelitic rocks in the immediate area of the Rich Mountain ultramafic bodies include kyanite and garnet (Rankin et al., 1972) and are also indicative of the middle amphibolite facies of regional metamorphism.

Hearn et al., (1977a) reported quartz diorite from the country rocks surrounding the ultramafics on Rich Mountain. This rock was never found as an outcrop, but only as float near body 1. The float was found on the top of a ridge near the ultramafic body, so presumably the source is in the immediate vicinity. Mineralogy of the rock is plagioclase (An_9-An_{18}) + hornblende + clinozoisite + minor quartz. Texturally the rock is allotriomorphic-granular with a weak foliation and shows considerable evidence of deformation. Plagioclase grains are bent and show undulatory extinction as do the grains of quartz. Both clinozoisite and sphene show reaction relations with the amphibole. This rock is probably best termed a metadiorite.

Rankin (et al., 1972) have mapped the amphibolites surrounding the ultramafics and correlated these rocks with the Ashe Formation. The massive to weakly layered amphibolites surrounding the ultramafic rocks are thought to represent metamorphosed mafic lavas (Rankin et al., 1972). Elsewhere in the Ashe Formation mica gneiss has been

Table 1. Modes of the amphibolites.

Sample Number*	% Amph	% Plag	% Epid	% Qtz	% Sph	% Cpx	% Opa	% Gar
1	58.2	14.1	12.7	4.2	2.8	8.0	---	---
3	72.6	19.9	0.8	3.0	2.0	---	3.7	---
4	76.5	8.0	8.4	1.0	3.4	---	---	2.7
7	67.5	22.8	---	3.7	0.7	---	---	5.3
8A+	61.9	22.4	---	4.1	0.9	---	---	10.7
8B+	66.1	19.9	7.5	4.4	1.3	0.8	---	---
10	66.5	25.0	0.2	7.1	1.0	---	0.2	---
11	66.5	17.9	9.9	4.1	1.4	---	0.2	---
12	58.8	10.5	3.7	10.3	3.4	---	2.1	11.2

Amph--hornblende, Plag--plagioclase, Epid--epidote, Qtz--quartz, Sph--sphene, Cpx--clinopyroxene, Opa--opaque, Gar--garnet.

*Sample locations given on Figure 2.

+A and B denote different handsamples from the same outcrop.

equated with a graywacke protolith (Rankin, 1975). Some authors believe the distribution of heterogeneous rock types in the Ashe Formation indicates the unit is a metamorphosed melange (Raymond, 1977; Swanson and Raymond, 1977).

Coarse-Grained Granitic Rocks

Lose blocks of pegmatite float are found in the area surrounding ultramafic body 1. None of the pegmatite was observed in outcrop, but the size of the blocks (greater than one meter) and their location (on ridge and hill tops) suggests the pegmatites come from near or within the ultramafic body. Some large block of quartz may be indicative of zoning in the pegmatites or at least the development of a quartz core.

ULTRAMAFIC ROCK

The Rich Mountain ultramafic bodies consist of two discrete masses separated by about 3.5 km of amphibolite (Figure 2). Callahan (et al., 1977) have suggested, based on the interpretation of geophysical and soil geochemistry data, that one of the ultramafic masses (body 1, Figure 2) is composite in nature and is much larger than the area suggested by surface outcrop. As mentioned previously, this mass of ultramafic rock is not well exposed and nothing was found during the geologic mapping to argue against the suggested composite nature of the mass. However, the distribution of surface flot argues against the large lateral extent of the ultramafic body suggested by these less direct geophysical and geochemical techniques and thus the composite

Table 2. Modes of the ultramafic rocks.

Percent	Body #1 Sample #1	2	3	4	5	6	7	8*	Body #2 1	2	3	4	5
FO	94.3	44.5	84.3	1.4	96.3	71.4	81.2	37.9	93.4	95.3	93.1	1.4	---
Opx	4.0	3.6	8.4	57.2	1.6	20.1	16.2	2.2	3.1	3.5	5.4	95.7	97.2
Opa	tr	0.9	0.2	tr	0.5	tr	tr	---	tr	---	tr	---	---
S	---	50.0	tr	---	tr	1.3	0.2	---	tr	tr	0.2	---	---
Ta	---	0.3	0.7	---	1.2	1.5	0.9	10.8	---	---	0.3	0.4	tr
Ch	1.3	0.7	0.9	---	0.4	3.8	1.3	---	3.3	1.2	0.3	tr	---
Tr	---	---	0.2	---	---	0.2	0.2	21.2	0.2	tr	0.7	2.5	2.8
Antho	---	---	---	---	---	---	---	27.9	---	---	---	---	---
M	---	tr	5.3	41.4	---	1.7	---	---	tr	---	---	---	tr
Texturally Stable Mineral Assemblages													
Fo+Ta+Antho								x					
Ta+Tr+Opx						x		x					x
Fo+Tr+Opx						x			x	x	x		
Fo+Ta+Tr								x			x		
Ta+Antho+Opx								x					
Fo+Antho+Opx								x					
Fo+Ta+Opx		x		x	x	x				x	x		
Fo+Ta+M		x			x								
Ta+Opx+M		x			x								
Fo+Opx+M		x	x		x								

Fo - olivine, Opx - orthopyroxene, Opa - opaque, S - serpentine, Ta - talc, Ch - chlorite, Tr - tremolite
Antho - anthophyllite, M - magnesite

*Float sample of anthophyllite schist presumed to be from contact zone between ultramafic body and country rock.

nature of this body is questionable.

Ultramafic is a descriptive term used to describe the dunite, harzburgite, orthopyroxene and serpentinite found in the Rich Mountain bodies. Dunite is a olivine-rich rock with minor accessory chromian spinel and less than 8 percent orthopyroxene. Harzburgite, is composed mainly of olivine with 18-22 percent orthopyroxene. Orthopyroxenite is composed almost entirely of orthopyroxene with less than 2 percent olivine. Nomenclature for the ultramafic rocks follows the recommendations of the subcommission on the Systematics of Igneous Rocks of the International Union of Geological Sciences (Streckeisen, 1973). Mineralogy of the ultramafic rocks (Table 2, Figure 3) suggests that there is not a continuum in the bulk composition, but rather three distinct bulk compositions represented in the ultramafic bodies of Rich Mountain.

Metamorphic minerals including serpentine, talc, tremolite, chlorite, anthophyllite and magnesite occur to varying degrees in the ultramafic rocks. The amounts of these minerals are highly variable (Table 2), but in every sample the original mineral assemblage can be recognized. Rather than choose an arbitrary limit to distinguish between serpentinites, metaultramafic rocks and fresh ultramafic, the terminology of the primary rocks will be used almost exclusively.

Structural features identified with the ultramafic masses include

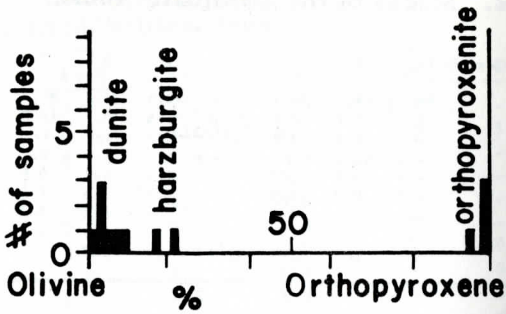

Figure 3. Histogram of rock types in the Rich Mountain ultramafic bodies.

joints filled with serpentine and magnetite, veins of talc, magnesite, anthophyllite and tremolite and discontinuous layers of orthopyroxenite. Surrounding the ultramafic masses is a metasomatic reaction zone representing a chemical interaction between the country rocks and the ultramafic rock. These reaction zones consist of tremolite-anthophyllite-talc schists. No attempt was made to collect structural data in the ultramafic bodies because of the relatively poor outcrops (body 1) and the discontinuous nature of the mineralogic layering (body 2). No systematic study was made of the vein mineralogy because Swanson (1979) has shown that the vein mineralogy is just a reflection of the mineralogy of the host ultramafic rock.

Dunite

Dunite is the most abundant rock type in the Rich Mountain ultramafic bodies and constitutes more than 90 percent of the observed outcrop area. Lithologically the rock is very uniform, with the primary assemblage olivine-orthopyroxene-chromian spinel. Serpentinization is best developed in the dunites and the color of the rock varies from light green to black as the degree of serpentinization increases. Grain size in the dunite varies from 0.1 to 2.0 mm. The olivine texture is xenoblastic-granular. Olivine in some samples forms a mosaic of polygonal grains that meet at 120° triple junctions, suggestive of an annealing fabric. None of the olivine shows any evidence of deformation (undulatory extinction or kink bands). Some of the samples with the texture suggestive of recrystallization or annealing show a foliation defined by a preferential elongation of some grains of olivine. This microfoliation was not apparent in the field. Orthopyroxene in the dunite is always slightly smaller than the associated olivine and is included within the olivine fabric as grains that do not show any evidence of deformation. Uncommon chromian spinel in the dunite usually forms subhedral to euhedral grains with a maximum size of 1.0 mm. Chlorite is associated with the chromian spinel. In some cases the chromian spinel grains are ragged and skeletal suggesting recrystallization.

Harzburgite

Harzburgite is only found in ultramafic body 1 on Rich Mountain where it comprises less than 10 percent of the rock mass. Contact relations between the dunite and harzburgite could not be determined in the field. At first appearance, it seems the only difference between the harzburgite and the dunite is an increase in the modal orthopyroxene content of the rock. Other features, such as olivine textures or mineral assemblages, do not vary significantly between the two rock types. However, the texture of the orthopyroxene is significantly different in the harzburgite. In addition to the orthopyroxene that is included within the olivine fabric and is very similar to that found in the dunite, the harzburgite contains porphyroblasts of orthopyroxene up to 5.0 mm long. The porphyroblasts have irregular grain boundaries and undulatory extinction. The large orthopyroxene grains are definitely not part of the recrystallization fabric that is dominate in the dunite and harzburgite.

Orthopyroxenite

Orthopyroxenites form discontinuous bands enclosed in dunite in ultramafic body 2 on Rich Mountain and comprises less than 10 percent of the outcrop area. Mineralogically the rock consists almost entirely of orthopyroxene with occasional olivine. Chromian spinel is not found in the orthopyroxenite. Tremolite, an uncommon phase in the dunite and harzburgite on Rich Mountain, composes 1-2 percent of the orthopyroxenite (Table 2). The bands of orthopyroxenite show evidence of deformation (Figure 4). Contacts between the dunite and included orthopyroxenite bands are very sharp and do not show any relations. Orthopyroxene in the orthopyroxenite shows an extreme range in grain size (0. 1 to 9. 0 mm). All of the pyroxenes have irregular grain boundaries and the result is a allotriomorphic-granular texture. The large orthopyroxene grains show undulatory extinction and kink bands while the smaller grains show little evidence of deformation. Tremolite is included within the orthopyroxene fabric as subhedral grains up to 1mm long and as small (0. 5 x 0. 1 mm) rod-like forms within the larger orthopyroxene grains. Tremolite found within the orthopyroxene is anhedral and is oriented parallel to the (210) pyroxene cleavage.

Metamorphic Mineralogy

Metamorphic minerals found within the ultramafic rocks include chlorite, tremolite, serpentine, talc, magnesite and anthophyllite. With the exception of serpentine, all of these minerals are found within the dunite in apparent textural equilibrium. Serpentine is found along fracture surfaces and is found replacing all of the Mg-silicates. The

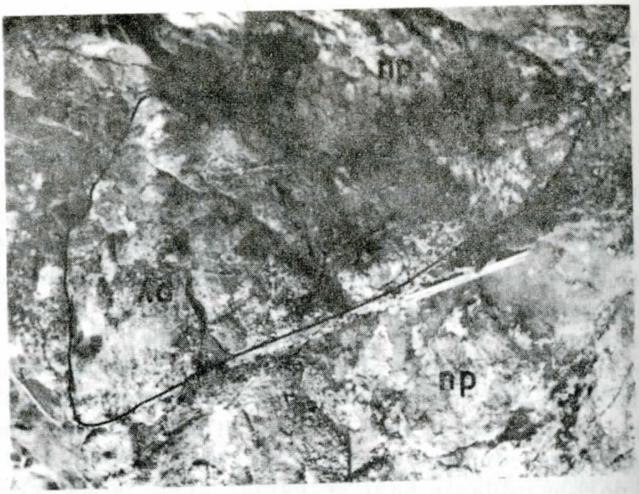

Figure 4. Outcrop of dunite (du) with included lens of orthopyroxenite (py). From body 2 on Figure 2. Scale is about 30 cm. in length.

overall abundance of the metamorphic minerals is low (Table 2) except for samples that show extensive serpentinization or samples from the metasomatic reaction zone between the ultramafic bodies and the country rock. Anthophyllite is found only within the metasomatic reaction zone surrounding the ultramafic bodies. Tremolite, although found in all the ultramafic rock types, is most abundant within the bands of orthopyroxenite.

Chlorite is perhaps the most ubiquitous metamorphic mineral in the Rich Mountain ultramafic rocks. Light brown, nonpleochroic chlorite is found in association with chromian spinel and as isolated subhedral flakes to 0.25 mm in diameter. Talc forms irregular grains up to 5.0 mm across and is in textural equilibrium with olivine and orthopyroxene. Fine-grained serpentine with a fibrous habit is common as both fracture coatings and in corona textures replacing the other Mg-silicates. Tremolite is present in the ultramafic fabric as subhedral to euhedral grains with a maximum length of 10 mm. This Ca-amphibole is also found as small rod-like grains included within and parallel to the orthopyroxene cleavage in the orthopyroxenite. Within the metasomatic reaction zone surrounding the ultramafic bodies, tremolite is found with a fibrous overgrowth of anthophyllite. Magnesite is found as anhedral grains in the dunite.

PETROGENESIS OF THE ULTRAMAFIC ROCKS

Primary Features

Bands of orthopyroxenite in the dunite of body 2 (Figure 4) are the only features that can be related to the initial intrusion of the ultramafic. Such masses of orthopyroxenite are common in alpine ultramafics (Loney et al., 1971) where they occur as dikes or sills within dunite. The deformed and disrupted nature of the orthopyroxenites on Rich Mountain does not facilitate identification of the original structures. Contacts between the orthopyroxenite and the dunite are very sharp on both the outcrop and thin section scale. No evidence of either contact metamorphism or chilled contacts has been found in association with the orthopyroxenites of Rich Mountain. Chen and Carpenter (1978) noted the occurrence of orthopyroxenite in the Frank (N. C.) ultramafic body, but they were unable to determine the relationship between the orthopyroxenite and the dunite.

Although the major minerals of the ultramafic rocks on Rich Mountain might be primary, some authors (Hearn et al., 1977a) argue that the present mineralogy is a product of regional metamorphism. Progressive metamorphism of ultramafic rocks will produce an anhydrous assemblage of olivine, pyroxenes and chromian spinel. Thus, it is not possible to attribute ultramafic mineral assemblages uniquely to either an igneous or metamorphic origin. The abundance of metamorphic minerals in the Rich Mountain ultramafic rocks (Table 2) indicates that these rocks have been metamorphosed and that the olivine and pyroxenes may be the products of prograde metamorphism of an ultramafic bulk composition such as a serpentinite (Carpenter and Phyfer, 1969; Hearn et al., 1977a). Tremolite intergrown with pyroxene in the orthopyroxenite bands suggests the former existence of a Ca-bearing phase (perhaps clinopyroxene) in the orthopyroxene that is now represented by the Ca-amphibole. Such textures strongly argue for a metamorphic origin for the orthopyroxene.

Some workers (e. g., Misra and Keller, 1978) have argued that the textures in ultramafic rocks such as those on Rich Mountain represent a primary crystallization event prior to emplacement of the rocks within the crust. However, others (Hearn et al., 1977a; Swanson, 1979) feel the textures are probably the result of recrystallization. The olivine fabric in the Rich Mountain ultramafic rocks, dominated by a polygonal arrangement of unstrained grains with straight grain boundaries that meet at angles of 120°, is a typical annealing texture and reflects recrystallization of the olivine (Spry, 1969). Within the orthopyroxenite, most of the pyroxene grains show evidence of deformation in the form of either kink bands or undulatory extinction. These textures suggest the olivine does not show any evidence of deformation while the pyroxene still retains kink bands and undulatory extinction. Recrystallization of the olivine has apparently removed any evidence of deformation while the pyroxene was not recrystallized and thus preserve a deformational fabric. Recrystallization of olivine and not pyroxene was also noted in the Frank ultramafic body (Carpenter and Chen, 1978). Experimental studies on the recrystallization of olivine and pyroxene in ultramafic rocks (Carter, 1971) show that enstatite recrystallizes at a higher temperature than olivine. Thus, a thermal event could have caused recrystallization of olivine and not pyroxene in the ultramafic rocks on Rich Mountain. Recrystallization textures of the olivines and the deformational fabric of the orthopyroxenes are clearly the result of metamorphism and are not primary magmatic features.

Metamorphism of the Ultramafic

Recrystallization textures together with texturally stable metamorphic mineral assemblages (Table 2) show that the ultramafic rocks on Rich Mountain have been metamorphosed. Preservation of sharp compositional boundaries, such as the contact between the dunites and the orthopyroxenites, suggests only limited ion mobility during the metamorphism. Indeed the ultramafic bulk compositions are themselves an argument for isochemical metamorphism, reflecting the metamorphism of an ultramafic protolith. The only components that appear to have been mobile during the metamorphism of the ultramafic rocks on Rich Mountain are H_2O and CO_2. The metasomatic contact zone surrounding the ultramafic masses is an obvious exception to the model of isochemical metamorphism since it represents a zone interaction between components from the ultramafic (chiefly MgO) and the country rock (mainly SiO_2, CaO, H_2O) formed in response to the extreme compositional difference between these rocks. Reoccurrence of simple, texturally stable mineral assemblages, such as olivine + talc + orthopyroxene (Table 2), suggests at least an approach to equilibrium during metamorphism of the ultramafic rocks on Rich Mountain.

Equilibrium phase relations involving the metamorphism of ultramafic bulk compositions can be treated in terms of the system MgO-SiO_2-H_2O-CO_2 and for assemblages involving a Ca-bearing phase such as tremolite, CaO-MgO-SiO_2-H_2O. Evans and Trommsdorff (1974) have developed a petrogenetic grid for the metamorphism of ultramafic rocks in the system MgO-SiO_2-H_2O-CO_2 based on the analysis of limited experimental work and the observation of natural assemblages. For the assemblages containing Ca-bearing phases, Evans (1977) has developed a sequence of mineral assemblages in the system CaO-MgO-SiO_2-H_2O and equated these assemblages with the facies of regional metamorphism. In all but the highest grade of metamorphism, Evans (1977) considered Al_2O_3 in the ultramafic bulk compositions to be buffered by the appearance of chlorite. The common occurrence of small amounts of iron in the ultramafic bulk compositions is thought to have little effect on the predicted phase relations (Evans and Trommsdorff, 1970; Winkler, 1976).

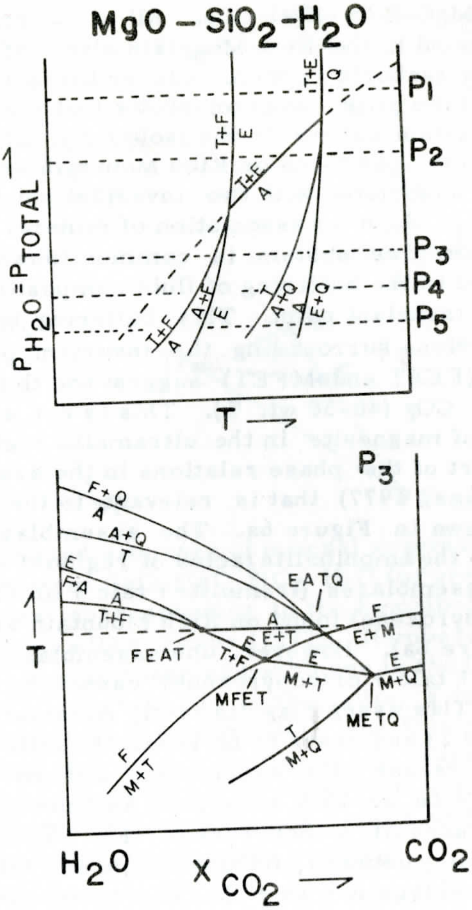

Figure 5. Phase relations in a portion of the system $MgO-SiO_2-H_2O-CO_2$ from Evans and Trommsdorff (1974). Rich Mountain assemblages (Table 2) include all of the univariant reactions involving invariant points FEAT and MFET. Phases shown include quartz (Q), talc (T), anthophyllite (A), orthopyroxene (E), olivine (F), and magnesite (M).

Evans and Trommsdorff (1974) discussed mineral assemblages in the system $MgO-SiO_2-H_2O-CO_2$ in terms of isobaric $T-X_{fluid}$ sections. Figure 5 shows a portion of the system that is most appropriate to the mineral assemblages in the Rich Mountain ultramafic rocks. The isobaric section P_3 most completely depicts the mineral assemblages in the system $MgO-SiO_2-H_2O-CO_2$. All of the tremolite-free mineral assemblages found in the Rich Mountain ultramafic rocks (Table 2) are found on the P_3 isobaric section. Closer inspection of the mineral assemblages and the phase diagram shows that the mineral assemblages represent univariant curves in the isobaric $T-X_{fluid}$ section shown on Figure 5. Assemblages found on Rich Mountain are represented by univariant curves associated with two invariant points (FEAT and MFET) on Figure 5, P_3. Such an association of mineral assemblages with invariant points in this system is common (Swanson, 1979) and is apparently related to the buffering of fluid compositions to values corresponding to the invariant point. This buffering is accomplished by the univariant reactions surrounding the invariant point. Both of the invariant points (FEAT and MFET) suggest the fluid contained a significant amount of CO_2 (40-50 wt. %). This is not surprising considering the abundance of magnesite in the ultramafic rocks on Rich Mountain.

That part of the phase relations in the system $CaO-MgO-SiO_2-H_2O$ (after Evans, 1977) that is relevant to the assemblages on Rich Mountain is shown in Figure 6a. The assemblages shown have been correlated with the amphibolite facies of regional metamorphism (Evans, 1977). Two assemblages (tremolite + talc + olivine and tremolite + olivine + orthopyroxene) found on Rich Mountain are depicted by tieline geometry (Figure 6a). However, one assemblage found on Rich Mountain (tremolite + talc + orthopyroxene) cannot be depicted by the geometry shown. This assemblage is fairly common in the rocks on Rich Mountain (Table 2) and there is no reason to believe the assemblage is unstable. The reason this assemblage does not show up on the geometry of Figure 6a is that a Ca-poor amphibole (A) appears prior to the first appearance of orthopyroxene (E). The Ca-poor amphibole plots on the tieline between orthopyroxene and talc (Tc) and thus precludes the assemblage A + Tc + E. The tremolite + talc + orthopyroxene assemblage was noted during a study of the Day Book dunite (Swanson, 1979). An alternative to the geometry presented by Evans (1977; Figure 6a) is shown as Figure 6b. In this sequence orthopyroxene is introduced prior to the introduction of Ca-poor amphibole and thus the assemblage A + Tc + E is geometrically possible. This possible change from the predicted phase relations of Evans (1977) may be related to the CO_2 concentration in the fluids that accompanied the metamorphism of the Rich Mountain ultramafic rocks.

Serpentinization is a late stage process in the Rich Mountain ultramafic rocks. Serpentine is found replacing the metamorphic minerals such as talc and tremolite as well as orthopyroxene and olivine. The serpentine is found as fracture fillings cutting the ultramafic rock and as coronas surrounding the Mg-silicates. Serpentinization is believed to occur in the presence of an H_2O-rich fluid (Johannes, 1969) at low temperatures and pressures (Barnes et al., 1969). These conditions cannot be correlated with other mineral assemblages in the ultramafic rocks or with any assemblages in the country rock.

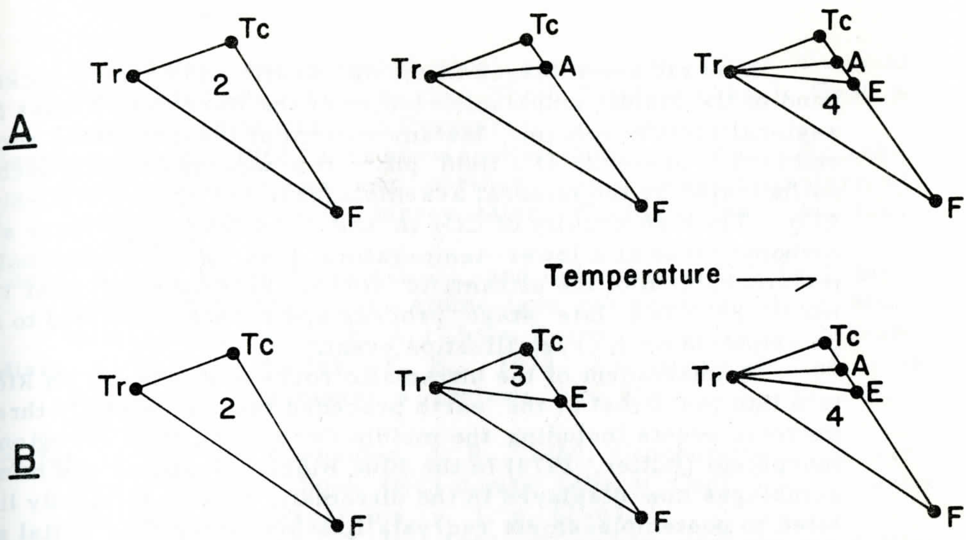

Figure 6. (A) Predicted sequence of mineral assemblages in a portion of the system CaO-MgO-SiO_2-H_2O for pressure-temperature conditions corresponding to the amphibolite facies of metamorphism, after Evans (1977). Numbers indicate observed occurrences of these assemblages in the Rich Mountain ultramafics. Tc is talc, Tr is tremolite, F is olivine, A is Ca-poor amphibole (anthophyllite), and E is orthopyroxene. (B) Predicted sequence of mineral assemblages in the presence of a moderate CO_2 concentration based on observed assemblages at Rich Mountain. See text for details. Abbreviations and numbers as in part (A).

CONCLUSIONS

Ultramafic rocks on Rich Mountain have been subjected to multiple periods of recrystallization during polyphase metamorphism as indicated by the mineralogy and textures. Textures in the ultramafic rocks suggest at least two and possibly three periods of deformation after crystallization of the ultramafic protolith. Olivine in the dunite and harzburgite shows textural evidence for a penetrative deformation that produced a microscopic foliation in the olivine fabric and a static recrystallization event that produced the polygonal olivine texture with 120° grain boundaries. Orthopyroxene is included within the fabric of these olivine-rich rocks. Within the harzburgite and orthopyroxenite are strained porphyroblasts of orthopyroxene. It is unclear whether these porphyroblasts represent a third recrystallization event, are associated with either or both of the recrystallization events that produced the olivine fabrics, or are relics from the original igneous rock.

Mineral assemblages in the ultramafic and country rocks correspond to the middle amphibolite facies of the Barrovian Facies Series of regional metamorphism. Metamorphism of the ultramafic rock occurred in the presence of a fluid phase that was moderately rich in CO_2, as indicated by the mineral assemblages in the system MgO-SiO_2-H_2O-CO_2. The high activity of CO_2 in the fluid phase may have stabilized orthopyroxene at a lower temperature than Ca-poor amphibole during metamorphism of the ultramafic rocks. Serpentinization of the ultramafic rock is a late stage process and cannot be related to any other deformation or recrystallization event.

Emplacement of the ultramafic rocks now exposed on Rich Mountain into the crust of the earth preceded two and possibly three metamorphic events including the middle Ordovician peak of regional metamorphism (Butler, 1973) in the Blue Ridge. Textures and mineral assemblages now displayed in the ultramafic rocks are mostly likely related to postemplacement recrystallization rather than initial emplacement into the crust. The original ultramafic protolith may have been a serpentinite (Phyfer and Carpenter, 1969; Hearn et al., 1977a) that was emplaced into a basaltic protolith, now represented by amphibolite. In any case, the original character of both the ultramafic and country rocks have been greatly changed during subsequent periods of metamorphism.

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POSSIBLE STRATIFORM-COPPER OCCURRENCE RAMSEYS DRAFT WILDERNESS STUDY AREA, AUGUSTA COUNTY, VIRGINIA

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ABSTRACT

Two vague northwest-trending belts of weakly anomalous copper values in stream sediment samples from Ramseys Draft Wilderness Study Area, Augusta County, Virginia, are probably related to low-grade stratiform copper occurrences in the Hampshire Formation of Late Devonian age. Soil samples from two of the anomalous drainage basins indicate a small mineralized area in greenish-gray sandstone on the divide between the drainage basins. Although no mineralized rock is exposed, the soil samples outline two copper-rich zones, 3-5 m thick. The copper content and mineralogy of the rock is unknown, but soil and rock samples suggest the possibility of a grade of as much as 1500 ppm copper. A strike length of 30 to 40 m and a downdip dimension of 45 to 55 m are consistent with the size of a mineralized area of that grade necessary to produce anomalous values in the two small drainage basins. Such small, low-grade stratiform deposits are not economically important, but this is the first such deposit found in the Upper Devonian red-bed sequence south of the known deposits in Pennsylvania. Additional prospecting for larger deposits in the extensive outcrop area of the Hampshire Formation in Virginia and West Virginia appears warranted.

INTRODUCTION

Stratiform copper deposits in red-bed sequences are known in the Catskill Formation of Devonian age in Pennsylvania (Weed, 1911, p. 59; McCauley, 1961; and Klemic, 1962). Stow (1955), Dennison (1973) and Dennison and Wheeler (1975) have speculated on the possibility that

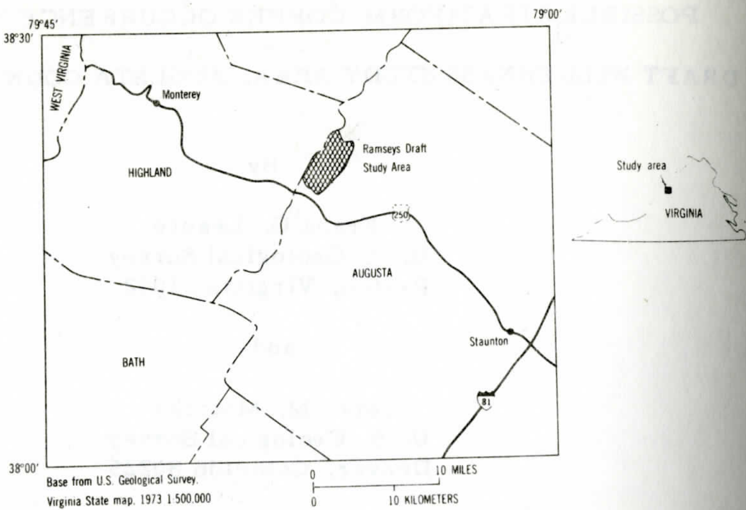

Figure 1. Index map showing location of Ramseys Draft Wilderness Study Area, Augusta County, Va.

similar deposits may be in the red beds of the Hampshire Formation, a correlative of the Catskill that is widely exposed in Virginia and West Virginia. During a mineral-resource evaluation of the Ramseys Draft Wilderness Study Area, Augusta County, Virginia (Figure 1), the geochemical reconnaissance based on stream-sediment sampling indicated two vague northwest-trending zones in the north half of the study area (Figure 2) in which anomalous values for copper were found (Lesure, Geraci, and others, 1977). In October, 1975 and 1976, we collected soil samples in the vicinity of these zones in order to evaluate the stream-sediment anomalies. Analysis of these samples has outlined a small area of copper mineralization in the upper Hampshire Formation along the divide between two small drainage basins. The results of this geochemical work are summarized here so that we may report the discovery of a stratiform-copper occurrence in the Devonian red-bed sequence south of the previously known deposits of Pennsylvania and also to illustrate a possible method for evaluating such discoveries.

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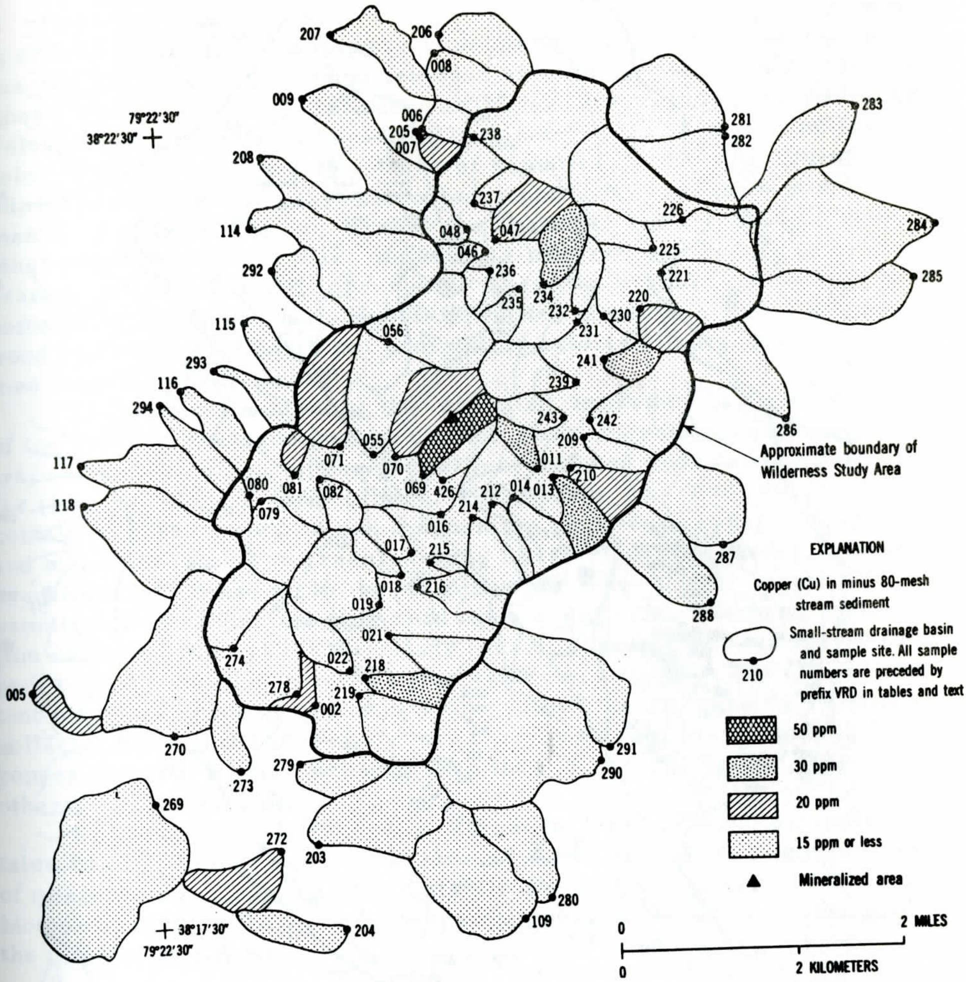

Figure 2. Map of Ramseys Draft Wilderness Study Area showing small drainage basins that contain anomalous copper values in stream sediment samples. Modified from Lesure, Geraci, and others, 1977, plate 3.

GEOLOGY

Ramseys Draft Wilderness Study area is a shallow syncline in Upper Devonian and Lower Mississippian sedimentary rocks (Lesure, Geraci, and others, 1977). Most of the area is underlain by the distinctive interbedded red and greenish-gray sandstone and shale of the Hampshire Formation of Late Devonian age. The overlying Pocono Formation of Early Mississippian age is preserved on higher elevations near the axis of the syncline. The underlying Jennings Formation is

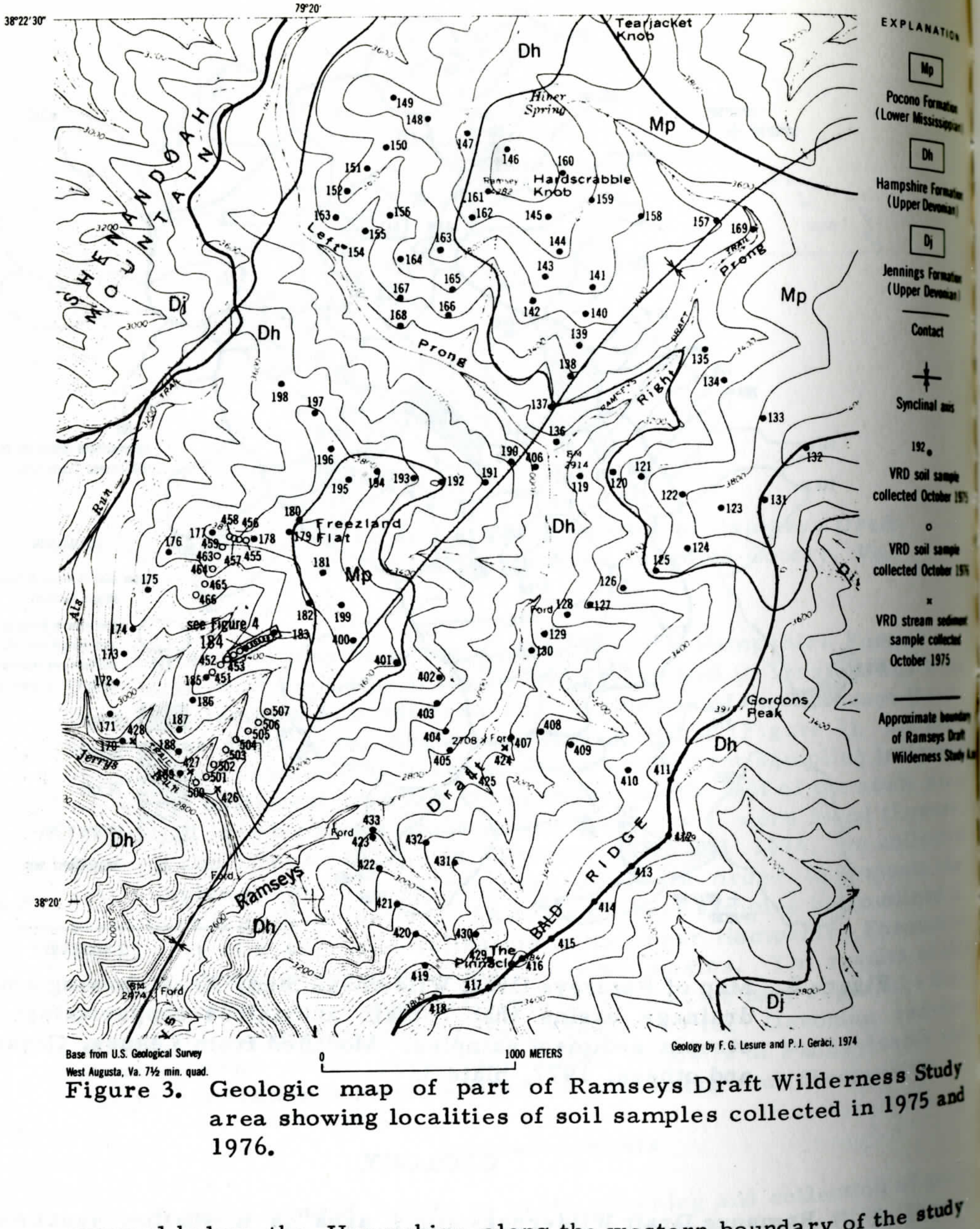

Figure 3. Geologic map of part of Ramseys Draft Wilderness Study area showing localities of soil samples collected in 1975 and 1976.

exposed below the Hampshire along the western boundary of the study area on Shenandoah Mountain (Figure 3).

The Hampshire is composed of thin- to thick-bedded, generally crossbedded, sandstone and interlayered mudstone. Most of the sandstone is grayish red to brownish gray; some is greenish gray. Crossbedded layers commonly 1-5 cm thick are in lenses 1-2 m thick. The sandstone is typically very fine to fine grained; a few beds are medium to coarse grained, and some are conglomeratic. The sand is primarily quartz, but minor amounts of feldspar and muscovite are also present. Chert, rock fragments, sericite, and clay are common accessories. Feldspar is generally altered to clay or sericite. The distinctive red color is produced by a fine coating of hematite on the sand grains. The interbedded greenish-gray sandstone commonly contains plant fragments, chiefly stems or grass-like material and irregular pieces of other plant parts. The plant material is now represented as thin, coaly seams, or iron-stained impressions where the organic matter has been completely removed. E. I. Robbins (written commun., 1977) has found wood cells and root hairs not completely oxidized in a sandstone specimen from Ramseys Draft.

The mineral content of the greenish sandstone is similar to that of the red, but the sand grains do not have hematite coatings. The trace-element content of the green-gray and red beds is also very similar (Lesure, Geraci, and others, 1977, table 2). The red beds may contain a little more iron and magnesium, but less copper, manganese, and zirconium than the greenish-gray beds. The differences, however, are probably too small to be significant. The copper content for 18 samples of the greenish-gray sandstone beds in the Hampshire Formation collected throughout the study area ranges from less than 5 to 70 ppm copper, and the median value is 5 ppm copper. The copper content for 17 samples of red sandstone from the Hampshire Formation collected throughout the study area ranges from less than 5 to 7 ppm copper and the median value is less than 5 ppm (Lesure, Geraci, and others, 1977, table 6).

One shale sample from the Ramseys Draft area (VRD 107) contains 70 ppm copper and has a trace of silver similar to that in analyses of mineralized shale in the Catskill Formation (Lesure, Weis and Motooka, 1977). Shale beds were not seen, however, near the site of the mineralized zones.

GEOCHEMICAL SAMPLING

Stream Sediments

A geochemical reconnaissance of the Ramseys Draft area was made as part of the mineral-resource evaluation of the Wilderness study area (Lesure, Geraci, and others, 1977, p. C13-C31). Eighty-four stream sediment samples, 38 pairs of soil and forest litter and 48 rocks were collected for analysis. The median value for copper in the 84 stream-sediment samples from the study area and vicinity is 15 ppm Cu; the range is from less than 5 ppm to as much as 50 ppm Cu. Sample VRD 069 from the drainage east of the mineralized area contains 50 ppm Cu, the highest value for any stream sediment collected. Sample VRD 070 from the drainage west of the mineralized area

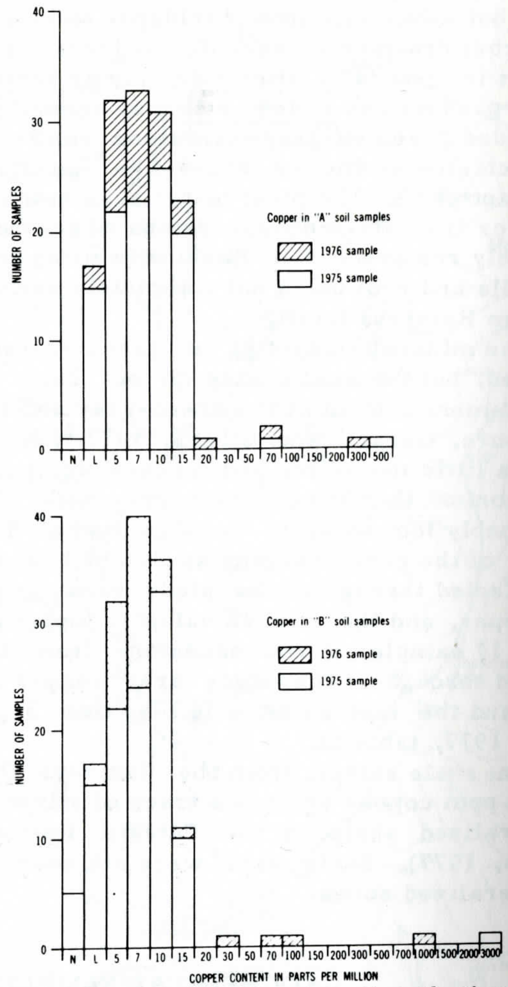

Figure 4. Histograms showing distribution of copper in "A" and "B" soil samples collected in 1975 and 1976. N, not detected; L, detected but below limit of determination.

contains 20 ppm Cu. The one high value of 50 ppm Cu might be considered slightly anomalous but not very significant. Values of 20 to 30 ppm Cu would not be considered anomalous except for the linear distribution pattern (Figure 2).

Table 1. Range and median values for 21 elements in soil samples from Ramseys Draft Wilderness Study Area, Augusta County, Virginia.

[All analyses are by a semiquantitative, six-step, D.C. arc, optical emission spectrographic method (Grimes and Marranzino, 1968). The values are reported as six steps per order of magnitude (1, 0.7, 0.5, 0.3, 0.2, 0.15, or multiples of 10 of these numbers) and are approximate geometric midpoints of the concentration ranges. Ca, Fe, Mg, and Ti values are in percent. All others are in parts per million. The precision is shown to be within one adjoining reporting interval 83 percent of the time and within two adjoining intervals on each side of the reported value 96 percent of the time (Motooka and Grimes, 1976). Symbols used: >, greater than value shown; <, less than value shown; N, not detected. Elements looked for but not found, except as noted, and their lower limits of determination (in parentheses): As(200), except sample VRD 184(700 ppm); Au(20); Bi(10); Cd(20); Mo(5); Sb(100); Sn(10); W(50); and Zn(200).]

	Soils on Hampshire Formation						Soils on Pocono Formation					
	Set A (105 samples)			Set B (109 samples)			Set A (37 samples)			Set B (37 samples)		
Elements (percent)	Low	High	Median	Low	High	Median	Low	High	Median	Low	High	Median
Ca	<0.05	0.5	<0.05	<0.05	0.07	<0.05	<0.05	0.3	<0.05	<0.05	0.05	<0.05
Fe	0.1	3	1	0.3	5	2	0.15	3	0.7	0.15	3	1.5
Mg	<0.02	0.5	0.1	0.05	0.5	0.15	<0.02	0.2	0.05	0.02	0.3	0.5
Ti	0.1	0.7	0.5	0.2	0.7	0.5	0.2	0.7	0.3	0.3	0.7	0.5
Elements (ppm)												
Ag	N	2	N	N	2	N	N	<0.5	N	N	1	N
B	10	100	50	15	200	70	20	70	30	30	70	50
Ba	50	700	200	70	1000	200	50	1000	150	50	700	200
Be	N	3	<1	N	5	<1	N	5	<1	N	2	<1
Co	N	20	<5	N	100	5	N	15	<5	N	20	5
Cr	<10	70	20	<10	70	30	<10	50	15	<10	50	20
Cu	N	500	7	<5	3000	7	<5	15	7	N	15	7
La	N	70	<20	N	50	20	N	50	20	N	50	20
Mn	20	>5000	150	10	>5000	70	20	5000	200	10	3000	200
Nb	N	70	<20	N	150	20	N	50	20	N	30	20
Ni	N	30	5	<5	70	7	<5	50	<5	<5	30	5
Pb	N	70	15	N	70	N	N	70	15	N	20	N
Sc	N	10	7	5	15	7	<5	10	7	5	10	7
Sr	N	100	N	N	150	N	N	100	N	N	<100	N
V	10	100	50	20	150	70	10	100	30	15	100	70
Y	<10	100	20	<10	100	30	10	150	20	15	50	20
Zr	70	>5000	500	70	>5000	500	100	700	300	150	700	300

Soil

In October 1975, we collected 219 soil samples from 110 sites in an area bounded by Hardscrabble Knob, Freezland Flat, Jerrys Run and the Pinnacle in order to evaluate the linear distribution pattern of the higher copper values in stream sediments (Figure 3). Two samples were collected at all but one sample site. Sample A was from the surface and contained abundant organic material. Sample B was taken below a sharp change in color and contained less organic material. The samples were dried in the laboratory; the A set was sieved to minus-150 mesh; the B set to minus-80 mesh. The samples were analyzed

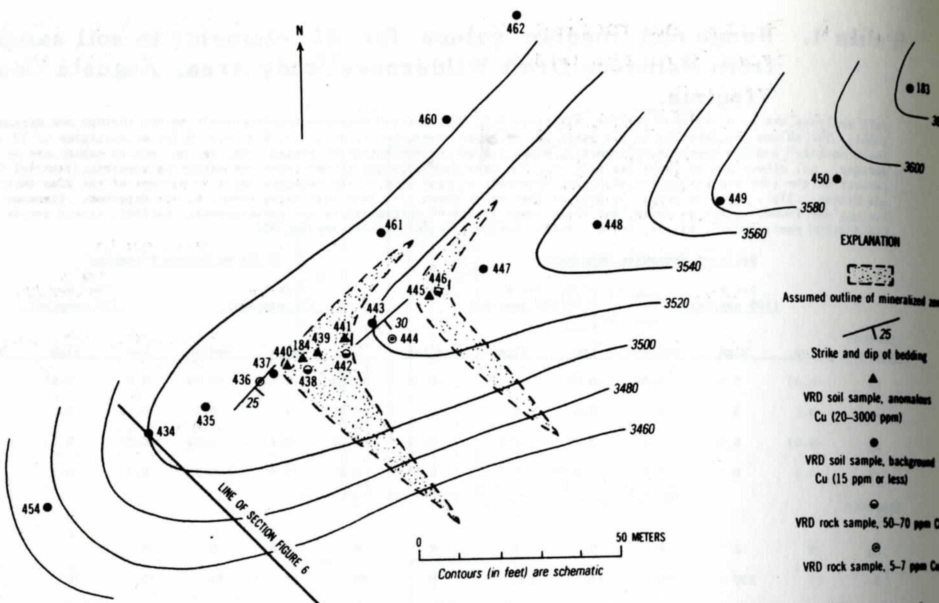

Figure 5. Sketch map of area of anomalous copper in soil samples showing sample localities and assumed outlines of mineralized zones.

for 30 elements by six-step, semiquantitative spectrographic methods in the Geological Survey laboratories, Denver, Colorado. A summary of the analytical data (Table 1) shows no significant differences between the two sets of samples, A and B, or between soils developed on Hampshire Formation and those developed on Pocono Formation.

All but two samples contained 15 ppm or less copper (Figure 4) similar to the soil samples collected in the earlier field work (Lesure, Geraci, and others, 1977, table 5). The two exceptions were collected at the same sample site, VRD 184. Sample A contains 500 ppm Cu and 2 ppm Ag; sample B contains 3000 ppm Cu, 2 ppm Ag, and 700 ppm As. No outcrops of mineralized rock were noted, but thin sections of weathered rock chips from the soil sample suggest that the bedrock is an arkosic sandstone that contained sulfide minerals(?) now altered to iron hydroxides. Additional analyses of a small composite sample of the weathered rock chips from soil sample VRD 184B, show 1500 ppm Cu, 1 ppm Ag, 300 ppm As and 150 ppm Co.

In October, 1976, we collected 69 more soil samples, 33 from 18 sample sites near the anomalous sample--VRD 184, (Figure 5) and the rest from spurs east and west of the anomalous site (Figure 3). These soil samples were collected in pairs, A and B, as before. We also collected three samples of greenish-gray sandstone and two samples of red sandstone from near the anomalous sample. Soil samples from four sites of this set, all within a few meters of sample VRD 184,

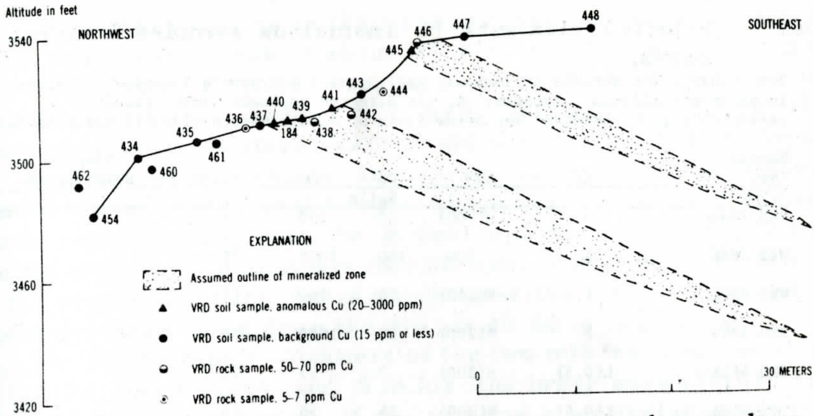

Figure 6. Generalized cross section in area of anomalous copper showing stratigraphic position of soil and rock samples and assumed shape of mineralized zones.

contain anomalous amounts of copper (Figure 6 and Table 2). The samples of greenish-gray rock, VRD 438, 442 and 446, also contained 50 to 70 ppm copper, more than the average for sandstone in the area. The samples of red sandstone (VRD 436 and 444) contain 5 and 7 ppm copper. Again, we found no obviously mineralized sandstone, but the association--copper and silver and an occasional trace of arsenic--is typical of stratabound copper deposits in Pennsylvania (Lesure, Motooka and Weis, 1977, table 1).

The area of copper mineralization is in the upper part of the Hampshire Formation on a southern spur of Freezland Flat (Figure 3). The area is on a less steep part of a narrow ridge containing scattered small outcrops of mostly red sandstone (Figure 5). Bedding in the sandstone strikes N. 30°-50° E. and dips 25°-30°SE. The east slope of the ridge is locally a dip slope on sandstone and is covered with a thin layer of sandstone debris. The west slope is thickly overgrown with mountain laurel and has few, if any, outcrops. Green sandstone is found mostly as float in areas between red sandstone exposures. There is nothing obviously different about this stretch of ridge as compared with other areas underlain by Hampshire Formation.

RESOURCE CALCULATIONS

It may be possible to estimate the size of a poorly exposed mineralized body of rock that will produce the anomalous stream-sediment samples found in the original survey. In a recent article, H. E. Hawkes (1976) discusses the downstream dilution of stream-sediment anomalies.

Table 2. Selected elements in anomalous samples from mineralized zones.

[see table 1 for details of analysis and figure 4 for sample locations: Values in parts per million. Symbols: N, not detected at lower limit given in parenthesis ; L, detected but below limit of determination given in parenthesis]

Sample No.	Ag	As	Co	Cu	Ni	Sample description
			Soils			
VRD 184A	2	N(200)	10	500	7	
VRD 184B	2	700	100	3000	70	
VRD 439A	1.5	N(200)	20	300	L(5)	
VRD 439B	2	N(200)	50	1000	30	
VRD 440A	L(0.5)	N(200)	7	20	20	
VRD 440B	L(0.5)	N(200)	15	30	20	
VRD 441A	1	N(200)	10	70	15	
VRD 441B	1.5	N(200)	15	100	20	
VRD 445B	L(0.5)	N(200)	15	70	20	
			Rock			
VRD 184	1	300	150	1500	50	Composite of rock chips from soil sample.
VRD 438	.5	N(200)	15	50	15	0.6m chip sample green-gray sandstone.
VRD 442	L(0.5)	N(200)	10	50	15	Composite of float blocks of greenish-gray sandstone.
VRD 446	1	N(200)	20	70	20	1-m chip sample of greenish-gray sandstone containing abundant plant impressions.

He gives the following idealized formula that relates metal content of an anomalous stream-sediment sample (Me_a) and the upstream drainage area (A_a) with the grade (Me_m) and surface area (A_m) of the mineralized area, provided the background (Me_b) of stream sediments is known and constant. The formula is:

$$Me_m A_m = A_a (Me_a - Me_b) + A_m Me_b$$

The eastern drainage basin has an area of 35,000 m^2 and stream-sediment anomalies(?) of 30 ppm copper (VRD 427) and 50 ppm copper (VRD 069). The western drainage basin has an area of 67,000 m^2 and an anomaly(?) of 20 ppm copper for both samples. The mineralized zones appear to be about 9 and 18 m wide along the ridge (Figure 5). We have assumed distances of 30 to 40 m along strike in the western drainage basin and 45 to 55 m down dip in the eastern drainage basin (Figures 5 and 6). The median value of 15 ppm copper in 84 stream sediment samples is used as background for stream sediments in the area.

Using these figures in the Hawkes formula, we estimate that an area of mineralized rock of about 700 m^2 in the eastern drainage basin would have a copper content ranging from 700 to 1700 ppm depending on which value of stream sediment is used. An area of 400 m^2 of mineralized rock in the western basin would have about 900 ppm copper.

Similarly if we assume an average value of 1000 ppm copper for the mineralized rock, on the basis of the average value of the five anomalous soil samples from the B zone and the analysis of rock chips from the soil, we can estimate that the area of mineralized rock extends into the western drainage basin along the strike for 20-40 m and into the eastern drainage basin down the dip for 45-60 m (Figure 5).

The stratigraphic thickness of the two mineralized zones is about 3 m for the upper zone and 5 m for the lower zone (Figure 6). The volume of mineralized rock, if we assume a lenticular shape for each zone, is about 4000 m^3. The specific gravity of several samples of greenish-gray sandstone from the Hampshire Formation ranges from 2.53 to 2.83 and averages 2.63. The tonnage of mineralized sandstone is therefore at least 10,000 metric tons and the copper-metal content might be as much as 10 metric tons.

It is also interesting to push the formula a little farther and estimate the order of magnitude of the copper content in an anomalous stream sediment indicating ore-grade rock. If we assume a copper content of 1 percent in a mineralized zone having the same size outcrop areas as were assumed previously, the anomalous stream sediment would contain about 215 ppm Cu in the eastern drainage area and 75 ppm Cu in the western drainage area.

CONCLUSIONS

The presence of a small stratiform-copper mineral occurrence in Ramseys Draft study area is suggested by the geochemical association of copper, minor silver, and a little arsenic in greenish-gray sandstone containing plant remains interlayered with barren red sandstone. The occurrence is probably too small to be of economic interest, but the possibility exists of larger and higher grade stratifrom deposits elsewhere in the Hampshire Formation. Areas of exposed Hampshire are large in western Virginia and eastern West Virginia. Our work suggests that stream-sediment and soil sampling are useful methods of exploration in looking for stratiform-copper deposits in this area.

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